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EVALUATION OF URANIUM ORE GUIDES, MONUMENT VALLEY, ARIZONA AND UTAH

CHARLES G. EVENSEN AND IRVING B. GRAY

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ABSTRACT

In Monument Valley of Arizona and Utah, on the southern end of the Monument upwarp, exposed sedimentary rocks range in age from Pennsylvanian to Jurassic, and are intruded by Pliocene (?) dikes and volcanic necks. Uranium occurs chiefly in and immediately below paleochannels cut into the Triassic Moenkopi formation and filled with sediments of the Shinarump member of the Triassic Chinle formation. Most ore occurs in local scours along these channel courses. Within the channels and scours,

uranium deposits appear to have been controlled largely by permeability. Marked increases in argillaceous content form impermeable barriers that localize a number of ore bodies. These barriers may be either argillaceous material within the Shinarump paleochannel sediments, or the steep channel flanks of Moenkopi mudstone. Carbonaceous material may have been a factor in development of an environment favorable to precipitation.

Primary uranium deposits in Monument Valley consisted largely of coffinite, uraninite, montroseite, and copper and iron sulphides. During oxidation, uranium and vanadium combine to form relatively insoluble carnotite-type minerals. Copper is migratory and generally is deposited in carbonates. In a few places copper-carbonates appear in halos surrounding uranium-vanadium ore bodies. Changes of color occur in the upper Moenkopi clastics in and around areas of more intense mineralization. Thickness of the zone of apparent color changes is extremely variable, apparently controlled by both intensity of mineralization and quantity of argillaceous material in superjacent Shinarump sediments.

INTRODUCTION

Scope.—The Monument Valley area of northeastern Arizona and southeastern Utah (Fig. 1) was examined by the U. S. Atomic Energy Commission and the U. S. Geological Survey as part of the search for uranium, and investigation of geology of its occurrence, on the Colorado Plateau. Field studies covered the entire Monument Valley district; special attention was given to the Oljeto area, where 200 holes were drilled through the Shinarump on Holiday Mesa, 94 on Flatiron Mesa, and 214 on Oljeto Mesa. This paper covers studies made by the Atomic Energy Commission during the period July 1953, to August, 1955. The authors are indebted to many associates in the Atomic Energy Commission for assistance in the studies and for critical review of the manuscript.

Geography.—The Monument Valley area, as referred to in this report, is a part of the Colorado Plateau, located on the Navajo Indian Reservation in San Juan County, Utah, and the northern portions of Apache and Navajo Counties, Arizona. It is bounded on the north by the San Juan River, on the east and south by Comb Ridge, and on the west by Nokai Canyon (Fig. 1). Monument Valley, in more restricted usage connotes a smaller area of spectacular erosional remnants along the Arizona-Utah state line between Comb Ridge and Utah State Highway No. 47. The surface is part of a dissected upland, most of which is between 4,500 feet and 6,500 feet above sea level. Routes of access to the area are shown on Figure 1.

Physiography of the Monument Valley area is characterized by steep-sided mesas, buttes and escarpments developed during an arid cycle of erosion and capped by resistant rocks. Drainage is by intermittent streams that are tributary to the permanent San Juan River to the north.

Previous Investigations and Mining.—Gregory (7) included Monument Valley in his regional study of the Navajo country, and the major portion of it was treated in detail by Baker (2). In 1951, the U. S. Atomic Energy Commission began its field investigations (3) and drilled small areas in the eastern, south central, and western portions. The U. S. Geological Survey mapped the geology of quadrangles in the Arizona portion during the summers

of 1951 and 1952, and carried out geologic strip mapping in the Utah portion during the summer of 1952. Grundy and Evensen, directed by E. D. McKee, studied the Shinarump conglomerate in the Monument Valley in the summer of 1952 for the Atomic Energy Commission (13). Following the close of Commission drilling on Hoskinnini Mesa in the fall of 1952, exploration for uranium by federal agencies was recessed. In the summer of 1953, the Commission began a study of ore controls and ore-finding criteria, which are the subject of this report.

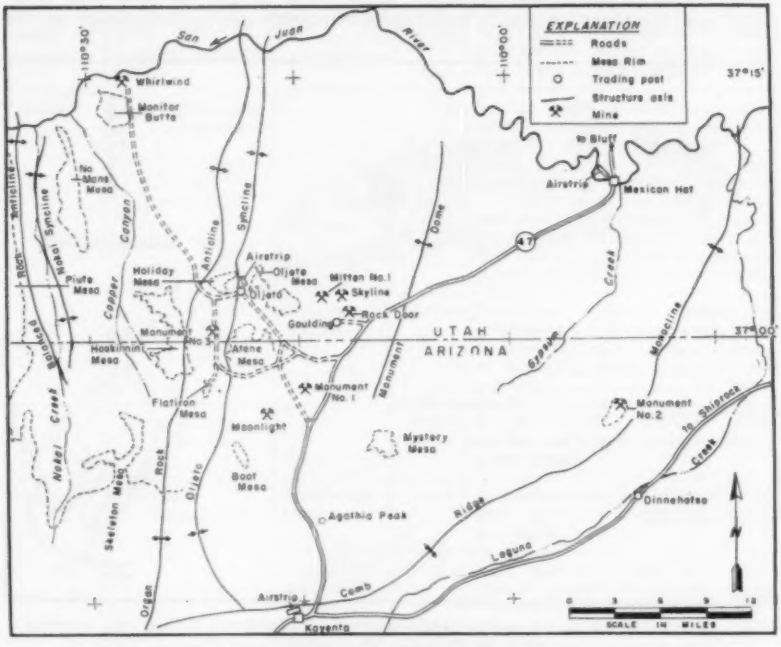

FIG. 1. Location map of Monument Valley area, Arizona and Utah.

The Vanadium Corporation of America, after obtaining leases on Monument Nos. 1, 2, and 3 mines in the late 1940's, was the first private operator of any consequence in the Monument Valley area. By 1952, the Whirlwind and Skyline mines (Fig. 1) had been opened and several small mines were operating in the vicinity of Monument No. 2. In the summer of 1953, when the Atomic Energy Commission resumed exploration in the Monument Valley area, only the Monument No. 2 was operating. During 1954, private exploration began which resulted in discovery of additional ore bodies. The trend toward deeper drilling has recently been successful in the Oljeto syncline.

STRATIGRAPHY AND STRUCTURE

Sedimentary rocks of Permian, Triassic, and Jurassic age crop out in Monument Valley (Table 1). They are almost entirely continental sediments of fluviatile and eolian origin. Basic dikes and volcanic necks, probably of Pliocene age, intrude these sedimentary rocks. Many sand dunes and a few gravel terraces conceal bedrock adjacent to the San Juan River and in tributary valleys.

Monument Valley is on the southern end of the Monument uplift, a north-south trending, broad, flat anticline. A series of subparallel secondary anticlines and synclines generally trending north-south are superimposed on the

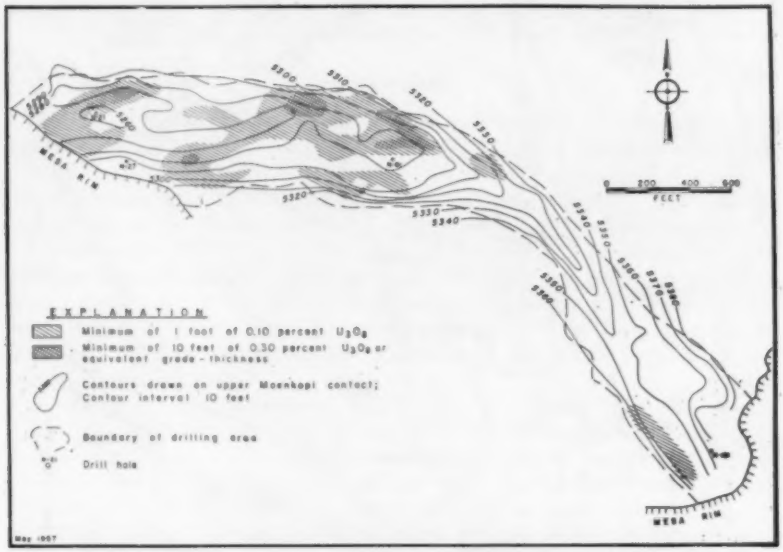
MESA RIM
EXPLANATION
Minimum of 1 foot of 0.10 percent U_3O_8
Minimum of 10 feet of 0.30 percent U_3O_8 or equivalent grade - thickness
Contours drawn on upper Moenkopi contact; Contour interval 10 feet
Boundary of drilling area
Drill hole
FEET

FIG. 2. Relationship of ore to configuration of Moenkopi-Shinarump contact before regional folding, Holiday Mesa, San Juan County, Utah.

major structure. From east to west across Monument Valley the secondary structures are Comb Ridge monocline, Monument dome, Oljeto syncline, Organ Rock anticline, Nokai syncline, and Balanced Rock anticline (Fig. 1). The anticlines are asymmetrical with dips averaging 10–20° to the east and 3° to the west. Erosion along the steeply dipping limbs has, in several areas, produced steep, east-facing escarpments parallel to the structural axes.

Jointing in the Monument Valley area is either transverse, oblique, or subparallel to the major structures. The most common directions are west, northeast, and north. Paleochannel sandstone of the Shinarump member appears to be less fractured than the thinner interchannel sediments (14, p. 174). Minor tension faults and bedding plane thrusts are associated with

TABLE 1

SEDIMENTARY ROCKS EXPOSED IN THE MONUMENT VALLEY AREA ARIZONA
(after I. J. Witkind, U. S. Geol. Survey Bull. 1030-C)

Age	Stratigraphic unit			Thickness (feet)	Lithologic and topographic characteristics
	Formation		Member		
Jurassic		Navajo sandstone		665	Pink and buff massive eolian cross-bedded sandstone with interbedded thin lenses of siliceous limestone.
Jurassic (?)	Glen Canyon Group	Kayenta formation		150–200	Light-violet to reddish-brown irregularly bedded quartz sandstone with local lenses of conglomerate.
Triassic		Wingate sandstone		350	Reddish-brown massive eolian cross-bedded sandstone weathering to steep rounded slopes or vertical cliffs.
	Chinle formation			1,000	Lower part, variegated mudstone and siltstone with dark sandstone lenses near base and thin beds of gray cherty limestone near top; upper part, reddish-brown even-bedded siltstone weathering to rubble-covered slopes.
			Shinarump conglomerate	50–100	Light-gray fluviatile cross-bedded conglomeratic quartz sandstone with rounded pebbles as much as 2 inches in diameter; much silicified wood present.
			—Unconformity—		
	Moenkopi formation			30–250	Reddish-brown even-bedded ripple-marked shaly siltstone weathering to gentle slopes.
Permian	Cutler formation		Hoskinnini tongue	10–60	Reddish-brown parallel-bedded siltstone with some interbedded lenses of sandstone.
			—Unconformity—		
			DeChelly sandstone member	350–450	Buff massive eolian cross-bedded quartz sandstone weathering to steep rounded slopes or vertical cliffs.
			Organ Rock tongue	650–750	Reddish-brown siltstone locally interbedded with red to gray sandstone; weathers to massive ledges about 5 feet thick.
			Cedar Mesa sandstone member	500	Reddish-orange cross-bedded sandstone with thin beds of gray limestone.
			Halgaito tongue	380	Red siltstone and silty shale with thin beds of nodular-weathering gray limestone.

regional structures. No evidence has been noted to indicate fracture control of primary mineralization, although secondary uranium, vanadium, and copper minerals are common along joints in and near the ore bodies.

Emplacement of ore was prior to regional deformation. Monument No. 2 and Moonlight ore bodies are now tilted, but correction for regional dip indicates that ore is localized in the bottoms of the scours as they existed before tilting (Figs. 2 and 3). Some evidence on Holiday Mesa, on the contrary, has been interpreted to indicate post-deformation mineralization, but the validity of the line of evidence used (distance that an arbitrary quantity of uranium occurs above the Moenkopi contact) depends on a uniform decrease

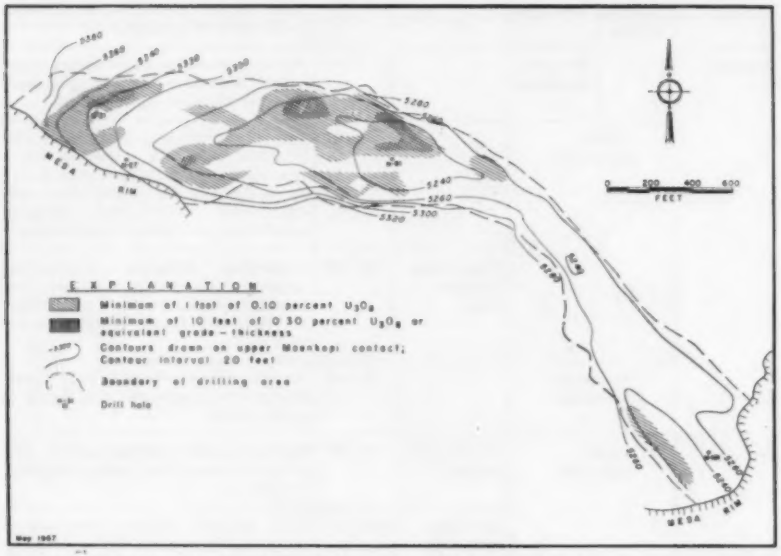

FIG. 3. Relationship of ore to configuration of Moenkopi-Shinarump contact after regional folding (i.e. at present time), Holiday Mesa, San Juan County, Utah.

in uranium content upward into the Shinarump, regardless of lithology. As is discussed later in connection with Shinarump lithology, this assumption is not generally valid. In any event, structural uplift followed by erosion has brought the Shinarump unit and its ore bodies near to the surface.

IGNEOUS ROCKS

The Monument Valley area is one of several on the Navajo and Hopi Reservations where dikes and volcanic necks of minette have been intruded into the sedimentary strata. The intrusive are considered to be of the same age as the Pliocene Hopi Buttes diatreme because of similarities in composition and structure (18). In most cases they appear to have followed north

to northwest fractures, such as the transverse fractures along the Comb Ridge monocline, according to Williams. Larsen and Short (2, p. 63) have described three intrusives as an augite minette, a hornblende gabbro, and lamprophyre.

All igneous rocks in the area are anomalously radioactive. Several dikes contain up to .02 percent uranium oxide and approximately equal amounts of vanadium and copper. Isotopic age determinations on uranium from Monument No. 2 mine indicate an age of 70 to 100 million years (Stieff, Stern, and Milkey, cited by Faul, 5, p. 264), and the beginning of the Pliocene is usually placed at approximately 10 million years. Thus the intrusives and uranium deposits in the Shinarump of Monument Valley are separated in time by 60 to 90 million years.

URANIUM DEPOSITS

Almost all uranium deposits of ore grade now recognized in the Monument Valley area are in paleostream channels, cut into the Triassic Moenkopi formation and filled with coarser sediments of the Shinarump member of the Triassic Chinle formation. In most places, ore extends into the subjacent Moenkopi formation and in one place, where channeling cut through the Moenkopi, into the Permian De Chelly sandstone. Concentrations of uranium minerals, possibly of ore-grade, have been noted also in the post-Shinarump Chinle beds at several places.

Uranium in the De Chelly Sandstone

The De Chelly sandstone tongue of the Cutler formation, of Permian age, consists of up to 450 feet of cross-stratified, fine-grained, eolian quartz sandstone. It forms vertical cliffs and rounded knobs. De Chelly sandstone underlies the Moenkopi formation, or the similar Hoskinnini tongue of the Cutler, over most of the area, but pinches out to the north, along Comb Ridge and a few miles south of the Whirlwind mine (Fig. 1); it is absent in the northern part of the area. At Monument No. 2 mine the paleochannel has been cut entirely through the Moenkopi formation and into the De Chelly sandstone. Here the De Chelly sandstone is mineralized to a depth of as much as seven feet with ore-grade uranium and associated vanadium. It has been suggested by Reinhardt (15) that this permeable floor below the channel was a factor in localizing the ore body.

Uranium in the Moenkopi Formation

The Moenkopi, oldest of Triassic formations in the Monument Valley area, consists of interbedded siltstones and mudstones. The Moenkopi and the underlying Hoskinnini tongue of the Cutler formation, which it was not practicable to separate during present studies, constitute a "red bed" unit which ranges up to 445 feet in thickness in Monument Valley. Monument Valley is near the extreme eastern limit of Moenkopi occurrence (12) and it has been entirely removed by erosion at Monument No. 2 mine.

The Moenkopi formation underlies the Shinarump member of the Triassic Chinle formation in almost all of Monument Valley. The erosional contact undulates moderately over much of the area and the contact zone is prominently bleached. In some localities relief of the contact may be as much as several hundred feet. McKee, Evensen, and Grundy (13) suggested that Moenkopi streams carved the surface on which Shinarump clastics rest and that the Moenkopi formation was unconsolidated at the time; additional erosional features appear to have been formed by early Shinarump streams.

Alteration at the Upper Moenkopi Contact.—The upper contact of the Moenkopi formation in Monument Valley is characterized by a zone of bleaching, as in a number of other areas on the Colorado Plateau. This zone is marked by the reduction and removal of ferric iron, and in the process the normal reddish-brown is changed to shades of yellow, green, gray, and white. The zone resembles in many respects the alteration zones in the upper parts of other red beds on the Plateau, where they are overlain by permeable formations. The alteration zone at the top of the Moenkopi formation is much thicker and more diversely colored, however, than the usual red-bed alteration, and it has been suggested that there may be a relationship between the thickness and color of the altered Moenkopi sediments and the quantity of composition of mineralizing solutions that passed through the permeable Shinarump above various portions of the contact. Observations on the thickness and color of the altered Moenkopi were recorded during rim examinations, mine mapping, and drill core logging, in order to test this hypothesis.

Alteration in the upper part of the Moenkopi formation is considered post-Shinarump in age, and not due to surface weathering or soil development prior to Shinarump deposition. This is indicated by its conformable relationship to the Moenkopi-Shinarump contact rather than to bedding within the Moenkopi, the extension of bleaching downward along fractures; the persistent occurrence of thicker alteration under more permeable Shinarump sediments; the predominance of white and light gray colors, which are generally rare as weathering effects; and the improbability that bleached rims and unaltered cores found in large blocks of Moenkopi material, within the Shinarump paleochannel-fill sediments, would be due to peripheral weathering of unconsolidated Moenkopi blocks which had never been compacted because of insufficient depth of burial. The most plausible explanation for the alteration appears to be the bleaching by ground water and/or mineralizing solutions percolating through the more permeable Shinarump sandstone.

The alteration zone at the top of the Moenkopi formation in Monument Valley varies in thickness from less than 1 inch to over 15 feet, averaging about 2 feet. The general prevalence of alteration throughout Monument Valley, and the Colorado Plateau as a whole, suggests that at least some of the alteration is a function of normal ground water movement. The reducing nature of the ground water may have been due in large part to the presence of organic acids and humic colloids derived from the abundant carbonaceous material in the overlying Shinarump and higher parts of the Chinle formation. In areas of uranium deposition, however, it appears likely that the mineralizing solutions, whatever their origin, were reducing in nature and that they have

caused additional alteration superimposed on the more widespread ground water alteration.

The depth of Moenkopi alteration in many places appears to be a function of the lithology of the Shinarump sediments directly overlying the contact. A comparison of the thickness of alteration to percent of mudstone in the lower 10 feet of Shinarump, from drilling data on Holiday Mesa (Fig. 4) and rim studies on Hoskinnini Mesa, indicates that the depth of alteration in many places is least beneath Shinarump sediments containing larger amounts of mudstone material. This is probably due to two factors: (1) the argillaceous material lowered the permeability of the Shinarump sediments; and (2) the

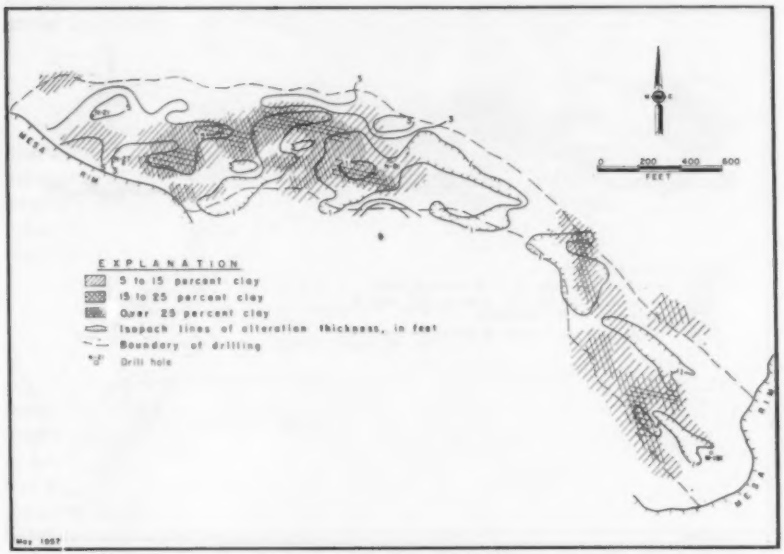

FIG. 4. Relationship of mudstone content in lowermost ten feet of Shinarump to alteration thickness in upper Moenkopi on Holiday Mesa, San Juan County, Utah.

mudstone material was derived from the Moenkopi formation and has also been altered by the solutions, thus probably lowering their reducing capacity.

Control of ore deposition by mudstone distribution in the basal portion of the Shinarump conglomerate is discussed in the section concerning the Shinarump member. The quantity of mudstone material appears to be especially important, and its relationship to both mineralization and alteration appears to be complex (compare Figs. 4 and 5).

Comparisons have been made between the colors of alteration in the Moenkopi rocks, immediately below the Shinarump contact, in areas of all variations from high-grade ore to barren rock along the Moenkopi-Shinarump contact. The colors in unoxidized (i.e. pyrite-bearing) and oxidized (pyrite altered to

limonite) areas differ, as might be expected, and comparisons were made separately for oxidized and unoxidized conditions.

Colors in oxidized areas cover a wide range of the color scale and, during mapping, a visual classification of the colors appeared adequate. Unoxidized colors, however, were confined to a more restricted color range and the National Research Council color chart was used for classification. This was done in an attempt to correlate minor changes in color with changes in intensity of mineralization. The same geologist made all the color comparisons. A good general correlation between alteration colors and intensity of mineralization appears to occur in both oxidized and unoxidized areas.

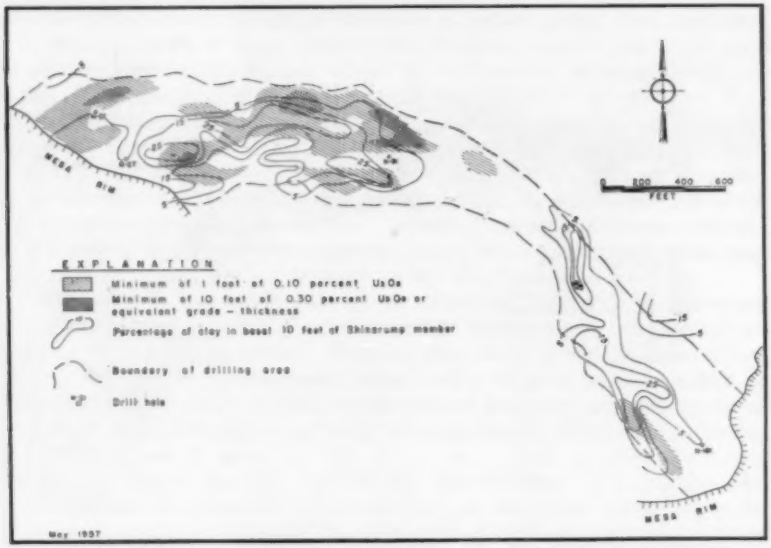
EXPLANATION
Minimum of 1 foot of 0.10 percent U_3O_8
Minimum of 10 feet of 0.30 percent U_3O_8 or equivalent grade – thickness
Percentage of clay in basal 10 feet of Shinarump member
Boundary of drilling area
Drill hole
MESA RIM
200 400 600
FEET
May 1957

FIG. 5. Relationship of ore to percent of clay in lowermost ten feet of Shinarump member, Holiday Mesa, San Juan County, Utah.

Colors of alteration in oxidized areas, in a general order of decreasing intensity of mineralization, are chalky white, gray, green, and yellowish-brown. No uranium is known to occur in unaltered reddish-brown mudstone, whereas the color in proximity to ore is generally chalky white to gray.

Two drilling areas, Holiday and Flatiron Mesas, are almost entirely unoxidized at the Moenkopi-Shinarump contact and the alteration colors were classified by comparison with the color chart. Because Holiday Mesa contains the first unoxidized ore bodies studied in this manner, no previous standard for color relationship to intensity of mineralization was available. Therefore, Holiday Mesa, where oxidization has extended only approximately 100 feet inward from the outcrop, was used as a standard for comparison with other

essentially unoxidized areas studied later. The various colors were tabulated with the average of their intensity of gamma radiation as an indication of the concentration of uranium associated with each alteration color. Selected color-groups were given arbitrary numbers, as listed in Table 2a and plotted in Figure 6. In contrast to favorable colors in oxidized areas, the grays and greenish grays are relatively unfavorable on Holiday Mesa whereas the yellow, brown and orange colors are more favorable. Data are not adequate to provide an explanation for this apparent reversal of the relation between colors and radiation intensity. Because these colors were ranked on the basis of relation to radiation in only one area, further verification in other areas would

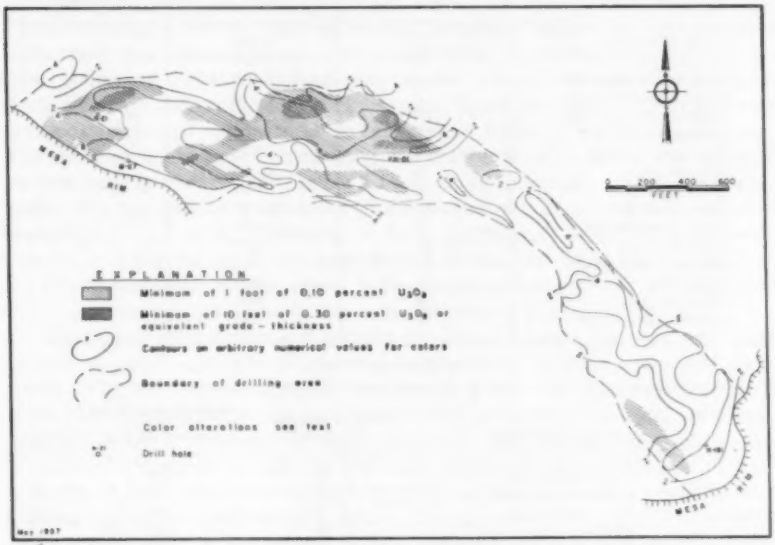
EXPLANATION
Minimum of 1 foot of 0.10 percent U_3O_8
Minimum of 10 feet of 0.30 percent U_3O_8 or equivalent grade - thickness
Contours on arbitrary numerical values for colors
Boundary of drilling area
Color alterations see text
Drill hole
0 200 400 600
FEET

FIG. 6. Relationship of ore to alteration colors, Holiday Mesa, San Juan County, Utah.

be necessary to prove the general applicability of the unoxidized color criteria. This study was continued in the unoxidized Flatiron Mesa drilling area and the results listed in Table 2b indicate that the only colors which occur here, below a channel containing only a small quantity of ore, are those found less favorable on Holiday Mesa. None of the eight most favorable colors in Table 2a occur in Table 2b. This appears to indicate a relationship between unoxidized alteration colors and intensity of mineralization, which should be tested further. The favorable colors on the east end of Holiday Mesa (Fig 6) may indicate the passage of uranium-bearing solutions but no ore was deposited in the relatively-flat-bottomed channel at that place. Further exploration in Monument Valley, involving drilling through greater thicknesses of

cover than is common at present, will almost certainly be in unoxidized material. This will afford further opportunities to test this suggested ore guide.

Gruner and others (8, p. 32) point out that under certain conditions ferric iron may be transferred by sulphate water into the crystal structure of the abundant hydromicas, forming gray to greenish-gray rocks. Only detailed

TABLE 2

COLOR CATEGORIES IN ORDER OF DECREASING INTENSITY OF GAMMA RADIATION

a. Holiday Mesa

Color of alteration	No. of samples	Average gamma log deflection (counts per sec.)	Color-group number
Dark olive gray	1	6,000	
Pale brown	1	6,000	5
Pale yellowish brown	6	5,300	
Light brownish gray	1	3,250	
Grayish orange pink	10	3,050	
Yellowish gray	17	3,000	4
Very pale orange	8	2,800	
Pinkish gray	4	2,750	
Very light gray	26	1,900	
Light olive gray	20	1,700	3
Light greenish gray	6	1,600	
Greenish gray	53	1,000	2
Medium light gray	13	600	
Light gray	22	600	1
Medium gray	4	400	
Medium dark gray	1	250	

b. Flatiron Mesa

Color of alteration	No. of samples examined	Average gamma log deflection (counts per sec.)
Light olive gray	19	950
Light bluish gray	2	750
Very light gray	3	750
Medium light gray	19	650
Light gray	10	600
Greenish gray	25	550
Light greenish gray	8	400

chemical studies would reveal the possible importance of this process and its relationship to uranium mineralization.

Uranium in the Shinarump Member of the Chinle Formation

All commercial uranium ore bodies in Monument Valley occur in the Shinarump member of the Triassic Chinle formation. This unit consists of lensing fluviatile deposits of light-gray to gray, cross-stratified, medium- to coarse-grained sandstone with a number of quartz pebble and mud pellet conglomerate lenses. Sandstone comprises approximately 75 percent of the Shinarump in Hoskinnini and Nokai Mesas (13, p. 12) and this figure is in

general representative for the remainder of Monument Valley. The Shinarump is resistant to erosion, forms steep cliffs, and caps many of the mesas and buttes in Monument Valley.

The Shinarump member varies from zero to 350 feet in thickness in Monument Valley. It is gradational both laterally and upward with other Chinle beds. The lower contact is a sharp erosional unconformity. The paleotopography of the Moenkopi-Shinarump contact is critical in the control of uranium ore and is discussed under the heading "Relation of uranium to basal contact and paleochannel lithology."

The Shinarump member pinches out along Comb Ridge between Monument No. 2 mine and the San Juan River, and also in the Clay Hills beyond the San Juan north of the Whirlwind mine. In those areas finer grained Chinle sediments rest directly on eroded Moenkopi beds. In both localities Shinarump-type sediments are interbedded with claystones of upper Chinle type, indicating that the pinchout may be one of facies change. Such transitional lithology does not occur on Holiday or Oljeto Mesas; the line of pinchout is some distance to the north in an area where Permian rocks now crop out. The southward pinchout occurs along the northeast side of Skeleton Mesa in an area largely covered with upper Chinle slump material. It has been suggested that the relative favorability of the Monument Valley area for uranium deposition is due to its proximity to these pinchouts. It is also noteworthy that the two highest-grade outcrops known in the post-Shinarump portion of the Chinle formation of Monument Valley occur in areas devoid of Shinarump (discussed under "Uranium in the post-Shinarump Chinle formation").

Shinarump sand grains and pebbles consist of quartz, quartzite, and chert as well as minor amounts of rounded fragments of igneous rocks and limestone. The argillaceous material consists of a few thin discontinuous mudstone lenses, mud pellets and mudstone blocks, common in channel sediments. Silicified wood is common throughout much of the member, ranging in size from small fragments to logs 50 feet long and five feet in diameter. Crossbedding, of both planar and trough types, is common, and the configurations indicate a general northwesterly flow of the streams that deposited the beds. Sedimentology of the Monument Valley Shinarup has been discussed in detail by McKee, Evensen, and Grundy (13).

Calcite, silica, and iron oxide are the cementing agents in the Shinarump member. Abnormally low amounts of calcium carbonate generally occur in unoxidized ore-grade specimens, and oxidation of the deposits, producing abundant sulphate ions, leads to further destruction of the calcium carbonate. "High-lime" ore is rare in Monument Valley. A zone high in silica cement, virtually an ortho-quartzite, occurs directly above high-grade ore at several places on Holiday Mesa. Its coincidence with more intense mineralization suggests that silica was a constituent of the mineralizing solutions.

Relation of Uranium to Basal Contact and Paleochannel Lithology.—The paleotopography of the Moenkopi-Shinarump contact is an important factor in the control of uranium ore deposition in the Shinarump member. Almost all ore in the Monument Valley area occurs in and immediately below paleostream channels cut into the Moenkopi formation and filled with Shinarump

sediments. Furthermore, almost all ore occurs in areas of scouring below the usual base of the channel (Figs. 2, 3). These scours resemble scours in modern streams and represent, topographically, the lowest areas on the Moenkopi-Shinarump contact.

Channels vary greatly in size, reaching a maximum width of approximately 3,000 feet and a maximum depth of 275 feet. In general, the largest channels occur along Copper Canyon, whereas "swarms" of smaller ones (averaging less than one-tenth as large) occur near Agathla Peak. Monument No. 2 mine, the largest ore body known in Monument Valley, occurs in a scour incised an additional 30 feet below the average channel depth of ten feet. The pattern of paleochannels resembles modern streams meandering within unconsolidated sediments, with numerous slumping channel banks (13). That Shinarump streams were apparently draining to the northwest is suggested by sedimentary structures within the Shinarump.

Channels outlined by Atomic Energy Commission drilling on Oljeto Mesa appear to be discontinuous, as described by Witkind (19). They are canoe-shaped with closed rather than open ends. It is suggested that these unusual features were formed by Shinarump streams that scoured downward through a thin veneer of unconsolidated Shinarump sediments into the fine-grained Moenkopi material below. Discontinuous channels appear to be rare in Monument Valley and it is possible that many channels identified as such are actually relatively deep scours along shallow channels (the scour at Monument No. 2 mine is considered an example).

Shinarump channel-filling clastic rocks, at or very near their contact with Moenkopi beds, appear to constitute the favored zone for ore deposition in Monument Valley (Fig. 7). If the quantity of uranium is small, it generally occurs in the upper foot of the Moenkopi formation. Where thickness and grade increase, most of the uranium is in the Shinarump, and richer ore in the Shinarump is accompanied by some increase in grade, and possibly thickness, of ore in the Moenkopi. In portions of Monument No. 2 mine, ore extends at least seven feet downward into the De Chelly sandstone. Several holes drilled on Holiday Mesa encountered ore in the Shinarump sediments some distance above the Moenkopi contact (Fig. 7). These appear to be instances of localized lithologic control.

Paleochannel sediments in general contain greater quantities of quartz pebble conglomerate, and carbonaceous and argillaceous materials, than the normal inter-channel Shinarump sandstone. This "trashy" Shinarump conglomeratic channel lithology appears to be especially favorable for uranium deposition in Monument Valley, as it does elsewhere on the Colorado Plateau. Clean sandstone is relatively unfavorable as a uranium host rock in Monument Valley and may be barren although in proximity to high-grade ore (Fig. 8). Carbonaceous matter, as is discussed under "Relation of uranium to fossil wood," appears to be an important precipitating agent. The role of argillaceous material is less well known, but in many Monument Valley localities it appears to be even more important than carbonaceous matter. Examples of apparent chemical precipitation or absorption of uranium by argillaceous material occur in several deposits, yet the rather general occurrence of small

quantities of apparently similar argillaceous material throughout the paleochannel sediments, where no appreciable amount of mineralization has taken place, appears to discount chemical and absorption effects to a large extent.

The physical effect of argillaceous material in forming impermeable barriers appears to be the most important control of ore deposition by that material. The lateral impermeable barriers may either be the banks of paleochannels where Shinarump sediments abut against less permeable Moenkopi sediments (Fig. 2), or highly argillaceous zones within the paleochannel sediments (Fig. 5). In relatively flat-bottomed channels, impermeable barriers formed by lateral increases in argillaceous content appear to be important ore controls. This is particularly evident in the relatively flat-bottomed channel on Holiday Mesa (Fig. 2), where the paleochannel itself is not as important

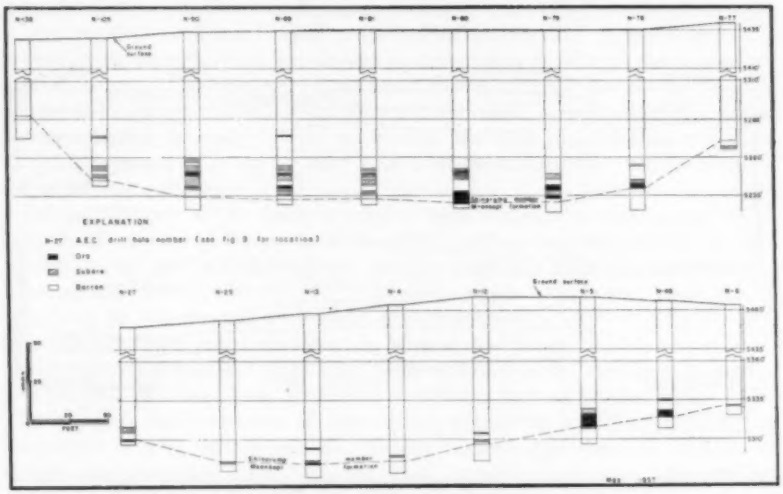

FIG. 7. Representative transverse sections, Holiday Mesa drilling area, San Juan County, Utah.

a control. Zones having steep channel banks as well as marked increases in argillaceous content are particularly favorable for uranium ore deposition, and ore bodies within the main Holiday Mesa scour appear to be controlled by at least one of these features. These relations may explain the almost universal occurrence of uranium in channels, i.e. they afford large areas of lateral contact between sediments of greatly differing permeabilities. If lateral changes in permeability are actually the primary control in uranium deposition in Monument Valley, other areas of similar lithologic changes within the Shinarump, outside paleochannels, would appear to be worthy of exploration.

Relation of Uranium to Fossil Wood.—Fossil wood, both silicified and carbonized, is common in the Shinarump member in Monument Valley. Silicified wood is more common in conglomerate lenses in the upper part of

the member, whereas carbonized wood is more common in the lower paleochannel sediments, although both may occur anywhere in the Shinarump or in overlying Chinle strata.

Uranium minerals coat outside surfaces and internal fractures of silicified wood, but in general they do not occur disseminated within the log. Thus, silicified logs are rarely ore grade, even though they may be surrounded by ore-grade sediments in many of the mines.

Carbonized plant material occurs in the Shinarump member as recognizable limbs and twigs, rarely as thin seams, and commonly disseminated in small flecks throughout sandstone and conglomerate lenses. Adjacent fragments of carbonized wood in the Skyline mine on Oljeto Mesa vary in uranium content by factors as high as 100. Although megascopic and microscopic study

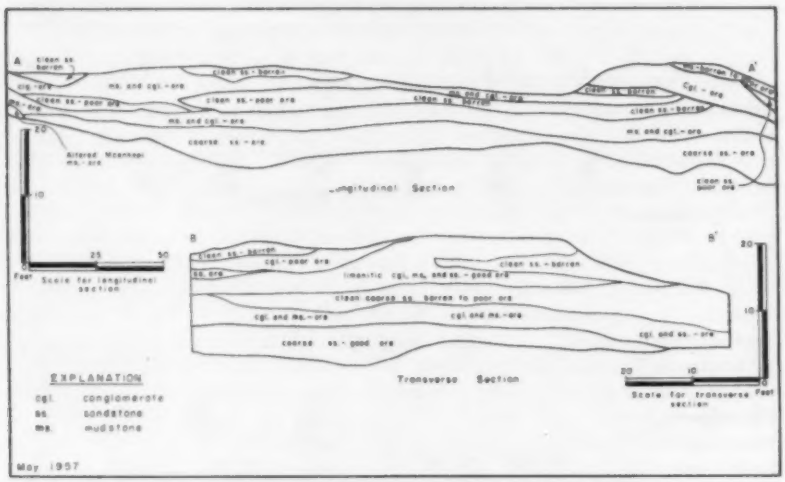

FIG. 8. Representative sections of the Monument No. 1 mine, Navajo County, Arizona.

of the material indicates no appreciable difference, some subtle variation in composition of the original wood or degree of carbonization may be the cause of the variation in uranium content. Detailed mapping of occurrences of carbonized material in mines and drill holes shows no direct correlation between carbonized material and ore, in many cases; however, greater concentrations of ore occur in scours that appear to contain more carbon, and possibly a certain amount is necessary for deposition of appreciable quantities of uranium. Pyrite occurs almost universally with "carbon trash" in unoxidized areas but appears to have no controlling influence on occurrence of ore.

Several logs with silicified cores and carbonized rims were noted during field investigations. It is suggested that these logs were partially decomposed prior to burial. The cells in the undecomposed cores were later replaced by

silica, perhaps derived from the destruction of volcanic glass in the overlying Chinle sediments, whereas the partially decomposed material eventually became carbonized. Perhaps the process is one of charring by forest fires, as has been suggested. This would require a considerable number of forest fires during late Triassic time as well as rapid burial to preserve the charred remains.

Horizontal Ore-Bearing Rods.—A number of horizontal "rods" (19, p. 119) containing high-grade ore occur in sandstone and conglomeratic sandstone lenses of the Shinarump in Monument No. 2 mine. These ore-bearing structures have been variously called rolls, logs, and rods. They differ from the better known characteristic rolls of the Morrison formation (16) in their general lack of recognizable bedding, their cylindrical shape, and relatively uniform cross-section over a considerable length. They consist generally of pipe-like forms up to 8 feet in diameter and more than 100 feet long, composed of structureless sandstone within cross-bedded sandstone. A few contain ill-defined cross stratification, not always corresponding to that in the surrounding sediments. The lithology of the sandstone in rods appears megascopically identical to the surrounding sandstones. Silicified wood fragments occur within a few of the rods, but no mechanism has been suggested for the complete replacement of a log by sand grains without disturbing the bedding in the overlying sediments.

Approximately 40 of these structures were mapped in Monument No. 2 mine by Evensen and Gregg in 1954. They are concentrated largely in four separate areas within the ore body and are generally aligned parallel to the trend of the channel (N. 18° W.) and at an angle to the other sedimentary structures in the channel that trend approximately N. 36° E. (13, p. 27). No rods have been noted elsewhere in Monument Valley.

Horizontal cylindrical pipes have been reported in the Arikaree formation of Nebraska and have been attributed to tubular flow of ground water (17). The physical characteristics of those pipes appear to be identical with the horizontal rods in Monument No. 2 mine, except that those in Monument Valley are cemented with uranium and vanadium minerals whereas the Nebraska pipes are cemented by calcite.

The Monument Valley rods might have formed by movements of subsurface water through the channel-fill sediments. Studies of cross-bedding indicate that Monument No. 2 channel was filled from the sides by Shinarump clastics (13). Tongues of these sediments would be saturated with water and, if at any time an outlet would open down the channel, much of the water of saturation would move laterally through the tongue parallel to the channel trend. Fraser (6) has pointed out that water moving laterally through a sediment tends to establish a passage of least total resistance. A passage with a circular cross-section would encounter the least resistance and, in an unconsolidated sediment, might destroy original bedding by grain-flotation. The occurrence of the rods in only four areas, totaling less than ten percent of the paleoscour area in Monument No. 2 mine, and not elsewhere in Monument Valley, indicates that the conditions that led to their formation and preservation were not widespread. Remnants of cross-bedding visible in several rods

suggest that flow of ground water was along the base of trough-type cross-strata sets; however, lack of general correlation between rod trends and sedimentary structures indicates that cross-strata were not an important controlling factor in water movement.

Upper Contact of the Shinarump Member.—Transition from Shinarump to Chinle lithology is characterized by gradation and intertonguing.

There is considerable evidence that the Moenkopi formation in the Monument Valley area was not compacted at the time of Shinarump deposition (13, p. 18). Later differential compaction would leave highs over Shinarump paleochannels, as a result of greater compaction of the bordering Moenkopi mudstones.

To check channel-tracing criteria, the upper surface of the Shinarump member was topographically surveyed in several areas where drilling was contemplated. On Little Oljeto and Oljeto Mesas, the surfaces are eroded to such an extent that no contour pattern is discernible, other than present drainage. On Holiday Mesa, however, only the western portion of the mesa has been appreciably dissected; a few mounds of residual post-Shinarump Chinle material remain in the center of the mesa. Following the topographic survey, elevations on the upper surface of the Shinarump were adjusted for regional dip and these adjusted elevations were contoured (Fig. 9). The axis of the channel, as determined by later drilling, has been superimposed on this map and appears to coincide rather closely with the trend of restored topographic highs. If this condition occurs on other mesas in Monument Valley, further topographic studies may be justified to delineate favorable ground for drilling. The rather special condition of erosion required for the topographic work, however, limits the method.

Geophysical and Geobotanical Studies.—Resistivity and seismic surveys were made on Holiday Mesa by the United States Geological Survey to ascertain the paleotopography of the upper Moenkopi surface. There the Shinarump member ranges from 50 to over 200 feet in thickness and varies markedly in lithology both vertically and laterally.

The resistivity studies encountered wide variations in surface resistivity and very low conductivity in the dry surface material of the mesa. Although Wayne H. Jackson (written communication, 1954) concluded that resistivity was not usable in determining the paleotopography under such adverse conditions, this was based on comparison with a projection of the channel, which is now known from drilling to have been in error. His interpretations do correspond in many respects to the channel as shown by later drilling.

The seismic investigation on Holiday Mesa (R. M. Hazelwood, personal communication) gave results that correspond very closely to the channel as outlined by later drilling. The bare rock surface and the unusual depth of the channel contributed to the accuracy of the survey. On the other hand the varying lithology of the paleochannel sediments caused some difficulty in determining an average velocity for the unit. It remains to be seen whether seismic studies will be as effective under less favorable conditions, such as areas of shallow channels or where appreciable amounts of Chinle material overlie the Shinarump member.

Indicator plants have proven to be of little value in the search for uranium ore in the Triassic sediments of Monument Valley. A few specimens of the selenium indicator plant *Astragalus* were found growing at random, and a few near Shinarump ore, but most were on surfaces of the Organ Rock tongue of the Cutler formation without any apparent relation to known ore bodies.

Geochemistry.—The ore deposits in the Shinarump conglomerate of Monument Valley are a complex of uranium, vanadium, and copper minerals emplaced in feldspathic, argillaceous sandstone and conglomerate. Considerable

EXPLANATION
Channel axis determined by drilling
Surface contours, corrected for structure
Rod stations
A.E.C. drill holes (See fig 7 for sections through drill holes)

FIG. 9. Surface topography corrected for structure, Holiday Mesa, San Juan County, Utah.

secondary silica was added in secondary growths on quartz grains and this, together with calcite, firmly cements the rock.

Early "rim" discoveries of uranium ore in Monument Valley were of carnotite type, commonly known as "yellow ore." During 1954 two relatively unoxidized "black ore" deposits were discovered, at Holiday Mesa and in the nearby Monument No. 1 mine. In these deposits oxidized ore occurs only near the surface, with a gradual transition from "yellow" to "black" ore over a distance of only a few feet. Gruner (10, p. 31) suggested that

yellow ores had been formed by oxidation of uraninite or coffinite deposits originally formed under reducing conditions. At least some of the "black ore" at Holiday Mesa and in the Monument No. 1 mine has been partly oxidized: valence of vanadium is the most sensitive index. Corvusite, the steel-blue mineral resulting from the partial oxidation of montroseite, is present in both.

The principal uranium and vanadium minerals in the "black ore" of the Monument Valley area are coffinite, uraninite, corvusite, and montroseite (E. B. Gross, personal communication). "Yellow ores" are a mixture of carnotite, tyuyamunite, metatorbernite, calciovolborthite, metarossite and hewettite (9, 11). Associated primary copper minerals are chalcopyrite, chalcocite, bornite, and native copper, and the secondary copper minerals, malachite and azurite, have been identified in hand specimens of oxidized ore.

It may be assumed that the relatively unoxidized ore bodies of Holiday Mesa and Monument No. 1 mine represent typical Monument Valley Shinarump ores before recent oxidation. Most of the ore bodies in Monument Valley, however, are of the oxidized "yellow ore" type and thus any geochemical generalizations must take both types into account.

From study of most of the oxidized deposits and the two relatively unoxidized deposits, both of which have oxidized portions, several generalizations are drawn. First, the uranium remains relatively fixed during and after oxidation in the deposits studied, because of an excess of vanadium over uranium. The ready combination of uranium with vanadium, of valence higher than in its primary occurrence, immobilizes the uranium in the relatively insoluble carnotite-type minerals. Where vanadium is available for such fixation of uranium, there is no significant migration of ore from the zone of primary deposition. Thus sedimentary and structural controls indicated by study of a mixed group of oxidized and unoxidized ore bodies will probably be valid ore-finding criteria for ores of both types. Second, copper is mobile during oxidation and is transported, probably by acidic solutions, until precipitation (generally as malachite) by the calcite cement in the rock. In a few cases, copper has combined with uranium to form torbernite.

The migration of copper ions and their deposition by the carbonate cement of the rock were investigated as a possible prospecting guide. Redeposition of secondary copper forms a copper carbonate zone surrounding the primary uranium-vanadium ore (Figs. 10, 11). Examination of other ore-bearing exposures in Monument Valley indicates that the conditions shown in Figure 11 are somewhat unusual. When the acidic solution carrying copper passed out of the ore body into the surrounding country rock, it encountered more than enough carbonate to neutralize the solution, and copper was precipitated, largely as malachite. Most ore bodies in Monument Valley, however, apparently contained larger quanties of acid-producing sulphides, such as pyrite and marcasite, and the carbonates were essentially destroyed for considerable distances from the ore body. This allowed the copper ions to disperse over a wider area, rendering them less valuable as a prospecting guide. This widespread migration of the copper, unfortunately, appears to be the general rule in Monument Valley. Copper concentrations may still be considered a useful

sign in prospecting, although they may be quite far removed from the original unoxidized ore.

Almost complete removal of copper from the oxidized rim ore body on Holiday Mesa (Fig. 11) indicates that if sufficient acid-producing sulfides are present, or if calcite cement is absent, copper might move considerable distances from the primary ore body. Workers in Monument Valley have pointed out an apparent copper zoning from a copper-deficient east side (Monument No. 2 mine of Comb Ridge) to a copper-rich west side (Copper Canyon). Assays of samples taken at the Moenkopi-Shinarump contact, down dip from Monument No. 2 mine, reveal up to .08 percent copper but

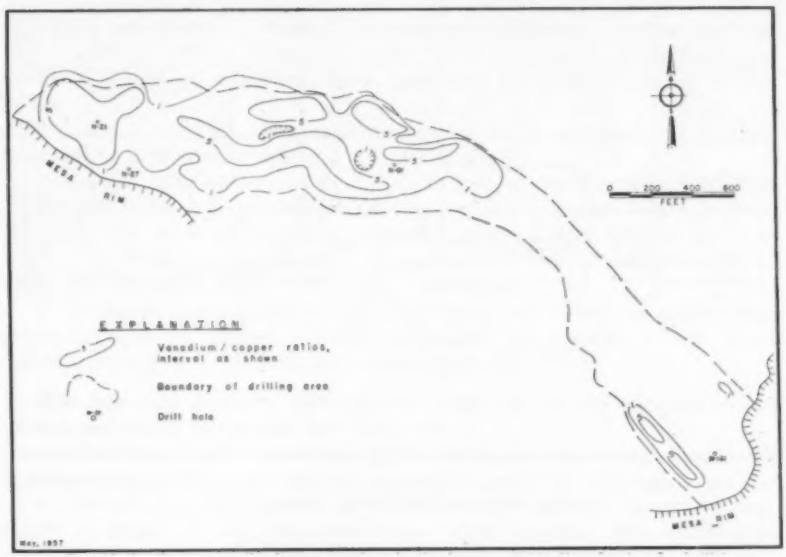

FIG. 10. Vanadium/copper ratios from asseys of uranium-bearing cores, Holiday Mesa, San Juan County, Utah.

no uranium or vanadium, whereas samples of ore from the mine rarely contain as much as a trace of copper. It is probable that the primary ore in the Comb Ridge area contained some copper, which was dispersed during oxidation.

An increase in the amount of vanadium eastward across Monument Valley, with practically none occurring on the west side, has been reported (1, p. 10). However, ratios of vanadium to uranium (averages taken from late production figures) for three mining districts in Monument Valley are as follows: Comb Ridge, 3.5:1; Oljeto, 4:1, and west side, 4.5:1. Thus, vanadium is present in all ores from Monument Valley, and variations in vanadium content within ore bodies appear to be more pronounced than area-wide variations.

Study of uranium, vanadium, and copper assays from drilling areas and

mine sampling shows no significant variation in the ratios of the three elements in unoxidized ore bodies. However, the higher vanadium-copper ratios are most common in areas of high uranium assays.

Spectroscopic analysis of ore, sub-ore, and barren rock samples taken during drilling on Holiday Mesa reveal the presence, in addition to vanadium and copper, of molybdenum, cobalt, chromium, and silver. Laboratory results from more detailed sampling will be necessary in order to test the possibility of using these elements as ore guides.

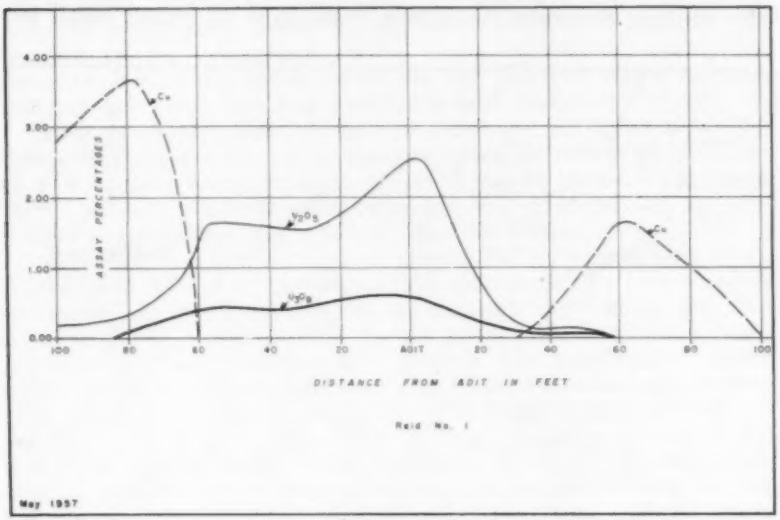

FIG. 11. Comparison of assays at ten-foot intervals along Moenkopi-Shinarump contact, west end of Holiday Mesa channel.

Uranium in the Post-Shinarump Chinle Formation

The post-Shinarump portion of the Chinle formation in Monument Valley consists of 800 to 1,000 feet of multi-colored lenses of mudstone and siltstone with scattered sandstone and limestone beds. Channeling and scouring are common throughout the sediments. Silicified and carbonized wood are abundant in portions of these upper Chinle units.

Two sub-commercial uranium occurrences are known in post-Shinarump portions of the Chinle formation of Monument Valley, in localities beyond the Shinarump "pinchout," where finer Chinle sediments rest directly on the Moenkopi erosion surface. One is on the northeast side of Skeleton Mesa in a carbonaceous sandstone lens of a paleochannel cut into Moenkopi. Several feet of Chinle mudstone separate the uranium-bearing lens from the Moenkopi contact. The richest sample from this prospect contained 0.30 percent uranium oxide for a thickness of one foot. The other near-commercial ura-

nium occurrence, drilled in 1954 by commercial interests, is along Comb Ridge between Monument No. 2 mine and the San Juan River. It is reported to be similar to the first.

SUMMARY OF ORE CONTROLS AND GUIDES

1. Almost all uranium ore in Monument Valley occurs in and immediately beneath paleochannels cut into the Moenkopi formation and filled with Shinarump sediments. Most ore occurs in scours along paleochannel courses. For reasons unknown, relatively deep scours along shallow channels appear to be the most favorable.

2. Reductions in permeability and impermeable barriers appear to be important ore controls. Marked increases of argillaceous content, such as clayey material within Shinarump paleochannel sediments or steep channel flanks of Moenkopi mudstone, have apparently localized a number of ore bodies.

3. Carbon, although not always ore-bearing itself, appears to be a factor favorable to precipitation within a channel or scour.

4. Certain color alterations in the upper part of the Moenkopi formation occur most commonly in and around areas of more intense mineralization. However, the depth of Moenkopi alteration is not a valid general guide to intensity of mineralization, inasmuch as argillaceous material within the Shinarump also influences depth of Moenkopi alteration.

5. Copper migrates during oxidation of Monument Valley ore bodies, until precipitated by carbonates. Copper carbonates are a guide to ore where distance of migration has been only a few hundreds of feet.

The following features, although not considered in the category of ore guides, are worthy of mention for future study:

1. Shinarump pinchouts may be significant factors in uranium distribution in Monument Valley.

2. Shinarump paleochannels, the localities most favorable for prospecting, might be traced by mapping topographic highs resulting from differential compaction, or areas of relative decrease in intensity of fracturing; both phenomena appear to characterize paleochannel sediments.

U. S. Atomic Energy Commission,
Grand Junction, Colorado,
Dec. 28, 1957

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GEOLOGIC CHARACTERISTICS OF FLUORSPAR DEPOSITS IN THE WESTERN UNITED STATES

W. C. PETERS

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ABSTRACT

A geologic study of fluorspar deposits in the Cordilleran region of the United States is presented in an effort to summarize mineralogical, structural and regional characteristics. Account is taken of the geochemistry of fluorine and comparison is made with fluorspar deposits in other areas.

In the geochemical cycle fluorine reaches marked concentration in the late stages of magmatic activity and in the biochemical-phosphatic stage of sedimentation. Though fluorine is easily removed from mineral combination, it is readily fixed in new environments before being dispersed. Thus, a region of high fluorine content should have a tendency to remain as such in spite of profound geologic changes.

Fluorspar deposits are greatly concentrated along the Eastern margin of the Cordillera in a pattern that may have existed through a large portion of geologic time, the present distribution being due in large part to areas of high fluorine content in the basement rocks.

Discordant fluorspar bodies have a tendency to be pipe-like and to display features that have resulted from collapse during mineralization. Textural features and paragenetic sequences in many Cordilleran fluorspar deposits indicate multiple stages of mineralization separated by stages of corrosion and

brecciation. Similar corrosion and collapse features are noted in non-fluorite ore deposits, but they are neither so widespread nor so striking as those that have involved fluorine-bearing hydrothermal solutions.

INTRODUCTION

General Statement.—In the Cordilleran region of the western United States, fluorite-bearing mineral bodies form a significant group of deposits with certain unique features.

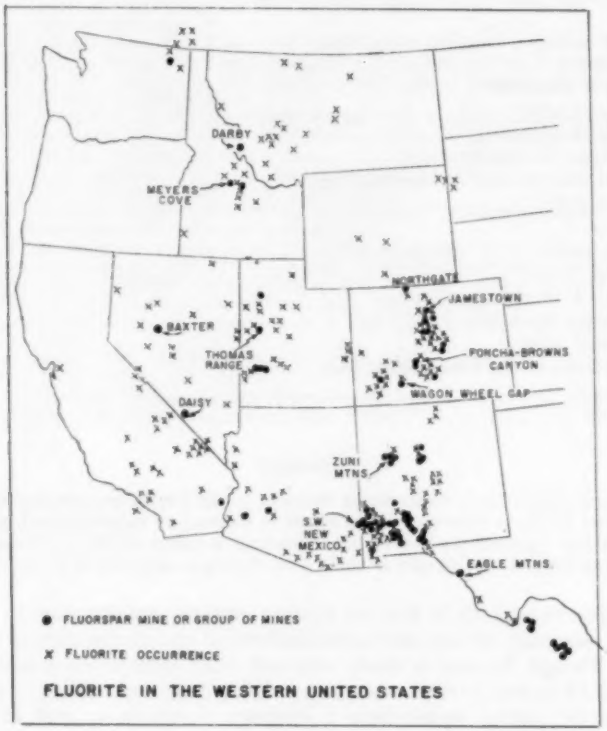


FIG. 1. Fluorite bearing deposits in the western U. S.

The most important industrial fluorspar districts within the region (Fig. 1) include: Deming-Silver City area, New Mexico; Zuni Mountains, New Mexico; Jamestown, Colorado; Northgate, Colorado; Salida-Poncha Pass area, Colorado; Thomas Range, Utah; Darby area, Montana.

Although the current market does not encourage intensive development of fluorspar deposits in the western United States, the long-range requirements for fluorspar justify continued investigation of distribution and genesis.

Field work on western fluorspar deposits was performed from 1952 until the present time in the course of the writer's employment with Westvaco Mineral Products Division, Food Machinery and Chemical Cororation. During 1955 and 1956, laboratory work was completed at the University of Colorado.

Previous Work.—The general economic aspects of western fluorspar deposits have been presented by Burchard (7), Gillson (16), Mudd (32), and Ladoo (25, p. 220–232). Lindgren (26, p. 152–180) has summarized the regional geologic pattern in the western states.

Among published works dealing with fluorspar deposits in specific states, the following are especially worthy of note: Cox (9), Colorado; Van Alstine (47), Colorado; Rothrock and others (35), New Mexico; Ross (34), Montana; Crosby and Hoffman (11), California; Thurston and others (45), Utah.

There have been published a large number of comprehensive studies on specific fluorspar districts. Outstanding contributions include those of Goddard (17, 18) on Jamestown, Colorado; Gillerman (14, 15) on the Eagle Mountains, Texas, and the Burro Mountains, New Mexico; Steven (42, 43) on Northgate, Colorado and St. Peter's Dome, Colorado; and Cox (10) on Meyers Cove, Idaho.

Acknowledgments.—The writer wishes to recognize the co-operation of numerous mine operators in releasing information to him and in permitting studies on their properties. During research under a fellowship at the University of Colorado, it was the writer's privilege to work under the guidance of Dr. L. A. Warner and Dr. E. E. Wahlstrom. The writer expresses his appreciation for their valuable criticism on the present manuscript and on the more extensive Ph.D. thesis from which it is taken.

Geochemical Considerations.—Significant contributions to the geochemistry of fluorine have been published by Barth (4), Koritnig (23), Borchert (5), and Goldschmidt (19, p. 568–583).

As a lithophile and biophile element, fluorine is typically associated with silicate rocks as a component of fluorapatite, of fluorite, and of certain complex silicates in which the hydroxyl ion (ionic radius 1.4 to 1.6Å) and oxygen (1.32Å) are subject to diadochic replacement. The ionic radius of fluorine, 1.33Å, allows significant quantities of the element to enter such hydroxl and oxygen-bearing minerals as mica, amphibole, tourmaline, and topaz.

In the process of magmatic differentiation, fluorine is a component of the volatile or "fugitive" phase. Early crystallization products such as peridotite, pyroxenite, and anorthosite are notably poor in fluorine. Apatite, the most important fluorine carrier, contains more chlorine than fluorine in mafic rock occurrences. The role of fluorine increases in significance during magmatic differentiation until it becomes a major constituent in granites and nepheline syenites due to large-scale substitution in hornblende and mica. The culmination of fluorine activity in the main course of magmatic differentiation is found in the development of pegmatite bodies and primary hydrothermal deposits.

The importance of fluorine in volcanism is readily seen from direct analyses of fumarolic gases. Some hot-spring waters of magmatic origin and of mixed magmatic-meteoric origin carry fluorine. Thermal springs at Poncha Springs,

Colorado, Wagon Wheel Gap, Colorado, and Ojo Caliente, New Mexico are of especial interest in that they are associated with fluorite-bearing veins.

Analyses of sedimentary rocks by Koritnig (23, p. 97) show that psammites and pelites are relatively rich in fluorine but carbonate rocks, cherts, and evaporites are relatively low in fluorine. The fluorine-bearing sedimentary minerals are micas, fluorapatites, clays, and topaz. Fine crystalline fluorite may also contribute to the fluorine content.

The mobility of fluorine during weathering and sedimentation has been discussed by Goldschmidt (19, p. 575) who suggests that under surface condi-

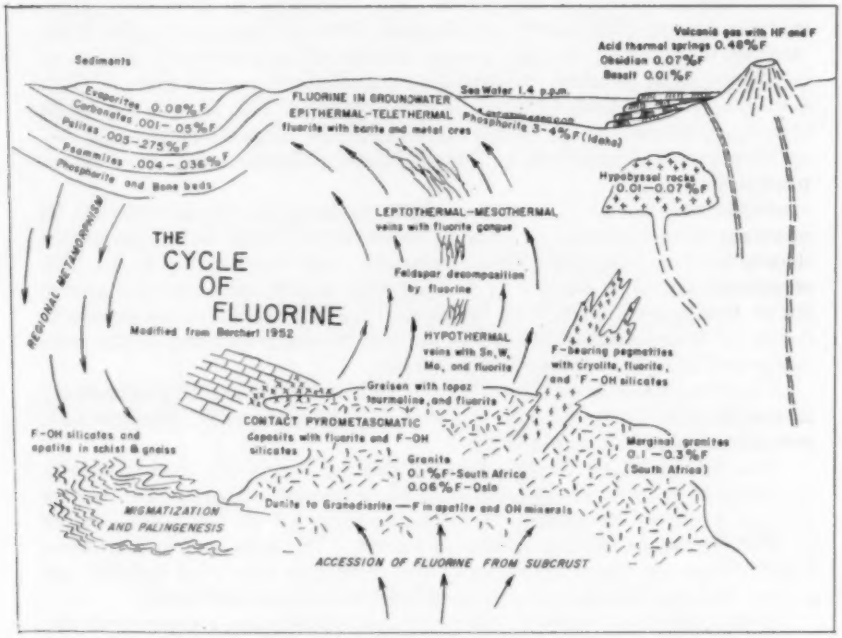


FIG. 2. Geochemical cycle of fluorine.

tions, there is some replacement of fluorine by the hydroxyl ion in micas and amphiboles. The writer has noted small veinlets of fluorite in phosphate rock in Idaho, thus illustrating a certain degree of mobility for fluorine that may have been leached from fluorapatite. A high measure of solubility for fluorine in surface water and groundwater is illustrated by areas in which fluorosis constitutes a health problem.

In marine waters, fluorine amounts to only a fraction of one percent of that contained in plutonic rocks. Barth (4, p. 421) comments that the relatively low concentration of fluorine compared to the large surplus of both chlorine

and bromine in sea waters serves to indicate extraction during transport. It is note-worthy in this respect that fluorapatite is actively forming along many continental shelves, thus removing fluorine from marine waters.

During metamorphism of sedimentary rocks, fluorine is made available for redistribution. Newly formed micas and amphiboles are probably fluorine-bearing. Granitic rocks derived from reconstitution of pre-existing sediments would display a fluorine content similar to that of magmatic granites.

The geochemical cycle, as outlined in Figure 2, involves appreciable amounts of fluorine in every stage, the average content of the lithosphere being, as suggested by Wassterstein (48, p. 211) and Barth (4, p. 420) approximately 0.08 percent by weight. The element reaches maximum concentration in the late stages of magmatic activity and again in the biochemical-phosphatic stage of sedimentation. During transport in surface waters, fluorine has a tendency toward rapid fixation in new sedimentary environments; thus, regions of high fluorine content would be expected to remain as such in spite of profound geologic changes. Significantly, regions in which there is an abundance of fluorspar deposits may have been regions of high fluorine content long prior to the time of hydrothermal mineralization. In the portion of the geochemical cycle that provides fluorine for mineral deposits, it would not seem necessary to seek derivation from the sub-crust. The regional association of fluorite with specific metallogenic areas may be a function of previous cycles of sedimentary differentiation followed by metamorphic and igneous differentiation that has taken place entirely within the crust.

Although fluorine is a constituent of a large number of minerals, the only fluorine mineral of commercial importance in the western United States is fluorite. The by-product recovery of fluorine from phosphorite during manufacture of superphosphate fertilizers has been proposed but is not yet a commercial reality.

CHARACTERISTICS OF WESTERN FLUORSPAR DEPOSITS

Morphology.—The characteristic features of fluorite-bearing deposits are not readily separable from those of ore deposits in general. Primary consideration will thus be given only to those deposits in which fluorite is either an economic mineral or in which fluorite represents a major factor in genesis. A morphologic division is made between concordant and discordant deposits.

McKinstry (31, p. 195–203) has summarized the features of concordant ore deposits; the major points in his presentation are applicable to concordant fluorspar bodies. He discusses the replaceability of limestone in contrast to other rock types, the effect of impermeable cap rock, and the relation between ore bodies and structural patterns. All of these features are expressed in the well-known "bedded" fluorspar bodies near Cave-in-Rock, Illinois. In the western states, fluorspar bodies in the Eagle Mountains, Texas, and near Challis, Idaho, are typical of concordant deposits.

In the Eagle Mountains, Texas, fluorspar occurs in concordant replacement bodies and also in fissure veins. As shown in Figure 3, a cross section of the largest replacement body in the district, the fluorspar zone is overlain by im-

permeable shales and by a gouge-filled fault zone. Within the mineralized zone, several beds of high grade fluorspar are separated by beds of shale, lenses of low-grade fluorspar, and bodies of unreplaced limestone. The replaced beds are believed by Gillerman (15, p. 62) to have been porous sandy limestones which were intercalated with the weakly mineralized shales.

A series of fluorspar prospects at Keystone Mountain, near Challis, Idaho, display many of the characteristics of concordant deposits. The mineralized zone is localized within a brecciated and silicified zone in dolomite. Immediately overlying the fluorspar zone is a hard silicified slate. The distribution of fluorspar exposures is shown in Figure 4. In addition to fluorspar bodies

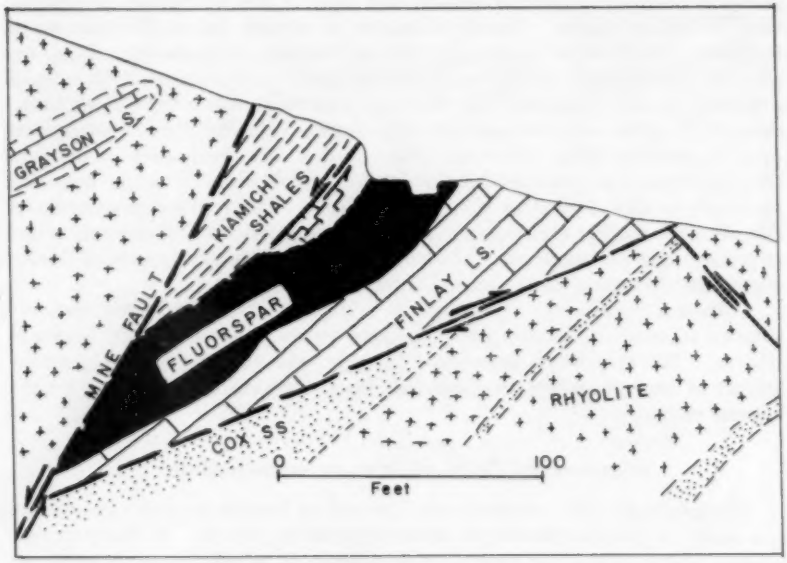


FIG. 3. Diagrammatic cross-section, north orebody, Eagle Mountains, Texas. After Gillerman (1953).

along the dolomite-slate contact, there are veins extending into the underlying dolomite beds along high angle faults. The largest fluorspar exposures lie near the junction of cross-faults with the stratigraphic contact. A second zone of discontinuous fluorspar exposures parallels the bedding but lies within the dolomite rather than along the slate contact. Inasmuch as there are several physically distinct phases of the dolomite and some interbedded zones of slate, it is probable that an unexposed slate caprock controls the lower mineralized zone.

The fluorspar bodies at Keystone Mountain lie along the flanks of a large anticline. The axial region of the anticline is brecciated and thoroughly silic-

ified but contains no fluorspar. According to Chambers (8), silicification in the axial area may have excluded the fluoritizing solutions.

Uranium mineralization in limestone near Grants, New Mexico, is closely associated with small bedded replacement deposits of fluorite. In some of the bodies, fluorite and calcite are the dominant minerals, with fine-grained uraninite incorporated in the fluorite crystals. The fluoritic uranium bodies lie within the Todilto limestone and locally within limestone lenses in the overlying shales of the Summerville formation. In areas in which the limestone is more than a few feet in thickness, the upper half of the formation is the favorable locus for mineralization. The upper half of the Todilto limestone

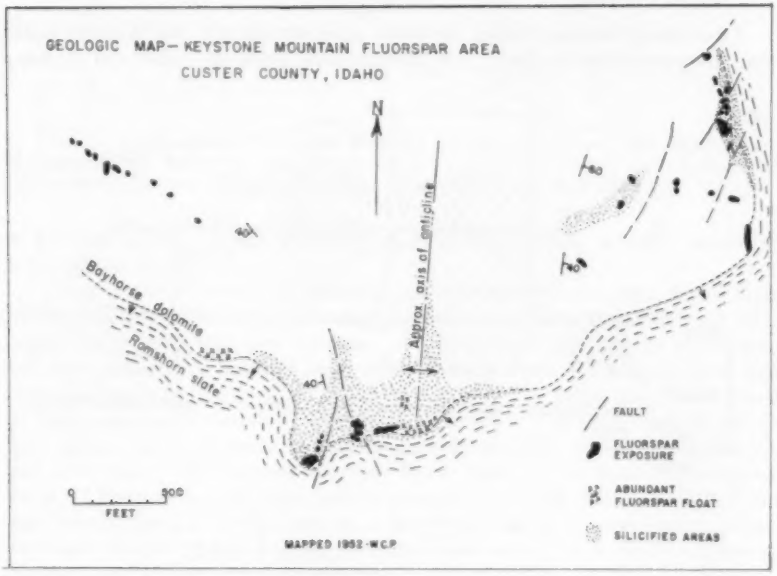


FIG. 4. Fluorspar area, Keystone Mountain, Idaho.

consists of seven to twelve feet of massive medium to fine-grained fragmental limestone with irregular zones of coarsely recrystallized calcite. The lower and less commonly mineralized part of the Todilto limestone is dark gray, fissile, and fine-grained.

The fluorite bodies within the Todilto limestone are irregular in outline. Some of the bodies have no apparent structural control; others exhibit a relation to the regional fold and fracture pattern. Some of the bodies have a tendency to follow the local joint pattern as mantos or as a linear series of tabular lenses. An association exists between broad gentle folds and some of the mineralized zones in that ore is most commonly found in crest or trough

areas. A more direct association is noted in the occurrence of sharply broken and crumpled synclines within some of the orebodies.

In summation, concordant fluorspar deposits are characterized by combinations of the following features:

1. Commonly in limestone, with a preference for open textured and coarse varieties.
2. Presence of relatively impermeable caprock.
3. Elongation of bodies related to fractures and resembling "manto" ore deposits.
4. Incomplete, erratic replacement within the mineralized zone, with lenses of unreplaced country rock.
5. Association with folds and collapse structures, as typified in the Cave-in-Rock district, Illinois.

Discordant fluorspar bodies are much more prevalent in the Western states than are concordant bodies. The characteristic types are veins and chimney

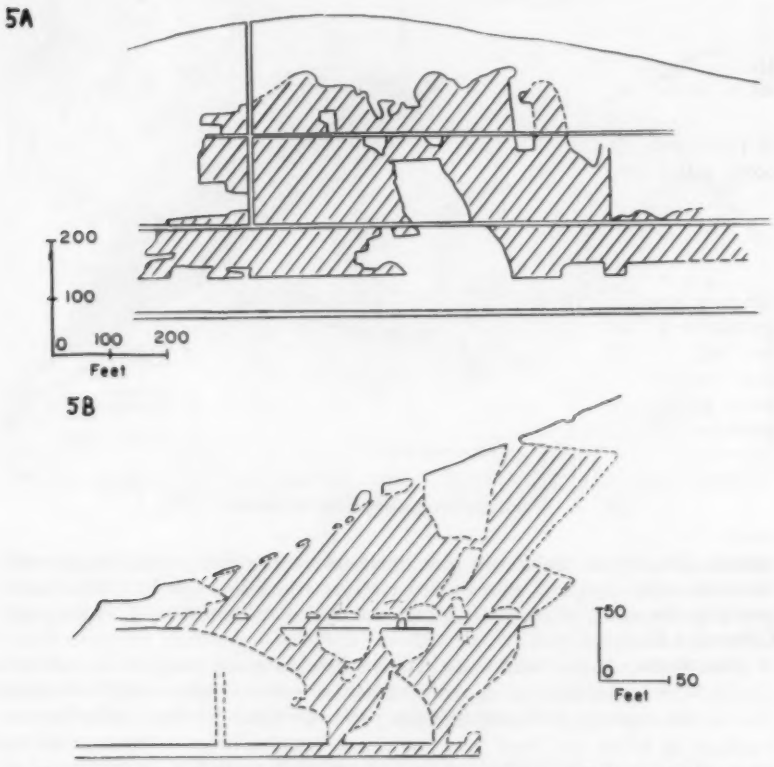


FIG. 5. (a) Longitudinal section showing stoped areas, Wagon Wheel Gap, Colorado. After Argall (3). (b) Longitudinal section showing stoped areas, Big Lead workings, Meyers Cove, Idaho. After Anderson (2).

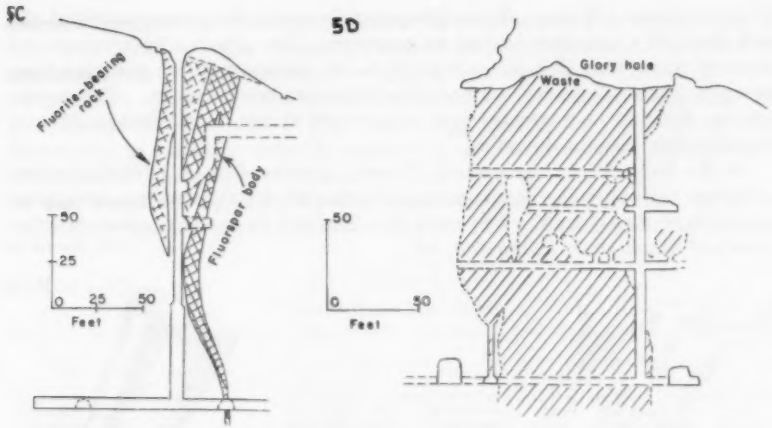


FIG. 5. (c) Cross-section, Old Hickory Prospect, Gallinas Mountains, New Mexico. After Rothrock and others (35). (d) Longitudinal section showing stoped areas, Emmett Mine, Jamestown, Colo. After Goddard (18).

or pipelike bodies. There is a marked tendency for steep pipelike shoots to occur within veins.

Veins in the Precambrian granitic and metamorphic complex of the Zuni Mountains, New Mexico, have many of the structural characteristics that typify discordant fluorspar deposits. A mineralized belt twenty miles long and three miles wide contains hundreds of steeply dipping fluorspar veins that range in length from a few feet to nearly a mile. The veins are characterized by relatively narrow width in proportion to length. The veins contain crustified and brecciated fluorite of several generations with minor barite, calcite, and iron oxide. High-grade fluorspar shoots occur, as in many metalliferous veins, at fracture intersections and in areas of relatively brittle rock. In the most productive part of the district, at the twenty-one and twenty-seven mines, fluorspar shoots plunge steeply and have been mined to a depth of several hundred feet with no apparent change.

Fluorspar shoots in vein systems display a tendency toward the development of longer vertical than horizontal axes in many districts. In many instances, the steeply plunging shoots have pipe-like characteristics. Finally, there are several prominent examples of pipe deposits. Examples of steeply plunging fluorspar shoots in vein deposits are found at Wagon Wheel Gap, Colorado; Meyers Cove, Idaho, in the Gallinas Mountains, New Mexico, and at Jamestown, Colorado (Fig. 5).

Fluorspar pipe deposits have been mined in the Thomas Range, Utah, and at the Daisy Mine, near Beatty, Nevada. Fluorspar in the Thomas Range occurs in a group of steeply plunging pipes, in smaller irregular bodies, and in a few small veins. The deposits are in massive fine-grained dolomite, which is underlain by a thick series of quartzite beds. Typical pipes in the Thomas Range (Fig. 6) are circular or elliptical in plan and range in diameter from

15 feet to about 130 feet. The pipes generally retain their cross-sectional size with depth to where they bottom on quartzite. One pipe has been mined to a depth of more than 200 feet. The pipes are associated with complex fracturing and, in many places, with rhyolitic intrusive breccia plugs. The contact between fluorspar and country rock is generally abrupt with a few small veinlets extending into the dolomite.

At the Daisy Mine, Nye County, Nevada, a series of pipes, irregular bodies, and veins with pipe-like shoots occur in a zone of chaotic structure associated with a large thrust fault. The pipes are localized in crackled zones in limestone and minor shale (Thurston, 44).

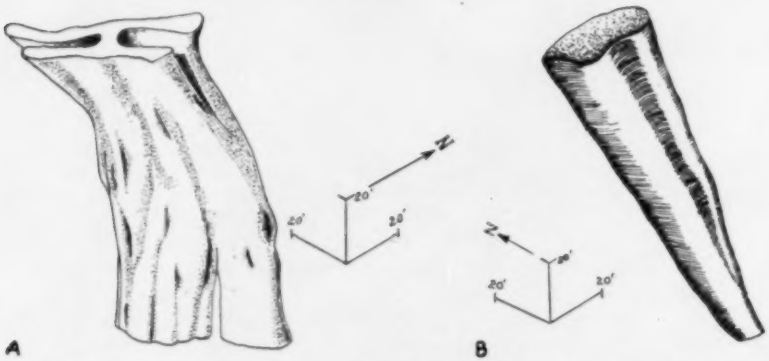


FIG. 6. (a) Block diagram, Bell Hill pipe, Thomas Range, Utah. After Staatz and Osterwald (40). (b) Block diagram, Lucky Louie pipe, Thomas Range, Utah. After Staatz and Osterwald (40).

In summarizing the characteristic morphology of discordant fluorspar deposits, two generalizations can be made:

1. There is a basic association between discordant deposits, their richer portions, and the fracture pattern of the wallrock. In this respect, fluorspar deposits do not differ from ore deposits in general.
2. There is a tendency toward chimney and pipe-like shoots in fluorspar. This feature is also in keeping with the character of ore deposits; it seems, however, to be especially pronounced in large fluorspar deposits.

Texture.—Textural features of commercial fluorspar bodies indicate great range in chemical and physical environment during their formation. Consideration is given in Table 1 to four basic textural types of fluorite and to two unique types, breccia texture and boxwork texture. The two latter types have widespread occurrence and special significance in indicating disturbances in the general sequence of deposition.

Major brecciated zones occur in a large number of Cordilleran fluorspar deposits. Minor brecciation occurs in nearly all deposits of commercial size. Brecciation appears to have taken place during mineralization, as shown by fragments of wallrock, fluorite, and associated minerals cemented by later

fluorite. Brecciation contemporaneous with mineralization is not confined to fluorspar deposits, but is especially common in such deposits. Brecciation in ore bodies has been attributed to mineralization stoping by Locke (27) and many others, and to collapse by McKinstry (31, p. 212–225), Joralemon (22, p. 256) and Goddard (18, p. 24). It is suggested in the present study that brecciation in fluorspar deposits commonly reflects alternating conditions of deposition and corrosion.

Brecciation during mineralization is apparent in a specimen from the Mirabal mine in the Zuni Mountains, New Mexico (Fig. 7); fragments of silicified

TABLE 1

TYPES OF FLUORSPAR TEXTURE

Textural Type	Typical Localities	Occurrence	Genetic Significance
Coarse-grained, massive	General	Anhedral to subhedral crystals in close integration. In pegmatites, pyrometasomatic assemblages, and hydrothermal open-space filling and replacement deposits	Not characteristic of specific conditions. May result from breccia or other types through secondary growth and recrystallization
Columnar	Wagon Wheel Gap, Colo. Northgate, Colo. Grantsville, Nev. Meyers Cove, Idaho	Broad to acicular columns, generally perpendicular to vein walls or radiating from fragments of wall-rock or of earlier generation fluorite. In open-space filling deposits	Concentrated source and quiet deposition enabling interfering crystals to grow outward from closely spaced nuclei. If radiating columns enclose fragments of fluorite, disturbance during mineralization is indicated
Coarse crystalline	General	Euhedral crystals, generally cubes. Crystals commonly zoned by color, unreplaced impurities, and liquid inclusions. In crustified veins, druses, and replacement deposits	Low concentration of fluorine source and slow nucleation as in weak or waning stages of mineralization or on the marginal area of a more concentrated body
Fine-grained, massive	Thomas Range, Utah	Compact aggregate of small crystals or crystal fragments. In open-space filling or replacement bodies	Complex and rapidly changing conditions. Concentrated source and rapid nucleation. In some cases, a fine breccia, indicating disturbance, or may be a "clastic" deposit from rapidly coursing fluids (Riedel, 1954, p. 250)
Breccia	Zuni Mountains, N. Mex. Jamestown, Colo.	Fluorite fragments and wallrock fragments cemented by later generation fluorite	Disturbance and collapse during mineralization
Boxwork	Keystone Mtn., Idaho Thomas Range, Utah	Residual mesh or network of fluorspar, associated minerals, and wallrock	Corrosion during mineralization. Intermediate stage to collapse breccia

granitic wallrock and of clear massive fluorite are cemented by a later generation of fine-grained fluorite. Another specimen from the Mirabal mine (Fig. 8) shows two stages of brecciation. Breccia fragments are cemented by fluorite, barite, and iron oxide; the fragments themselves are composed of brecciated coarse crystalline fluorite and granitic wall-rock in a matrix of finer grained fluorite.

Two stages of brecciation are apparent in a specimen from the Penber shaft, Northgate, Colorado (Fig. 9). Clear fluorite fragments are cemented by later impure dark fluorite; a fragment containing both generations is incorporated in a matrix of fine-grained fluorite.

In a large number of breccia specimens from various districts, secondary growth is apparent around the edges of fluorite fragments. In Figure 10 subhedral crystals of clear fluorite have partially replaced a fine-grained quartz-fluorite-hematite mass; successive boundaries from earlier stages of growth remain as "ghost" lines within the fluorite. With continued growth of the clear fluorite crystals, a coarse-grained massive texture would result, leaving very little indication of the original breccia texture.

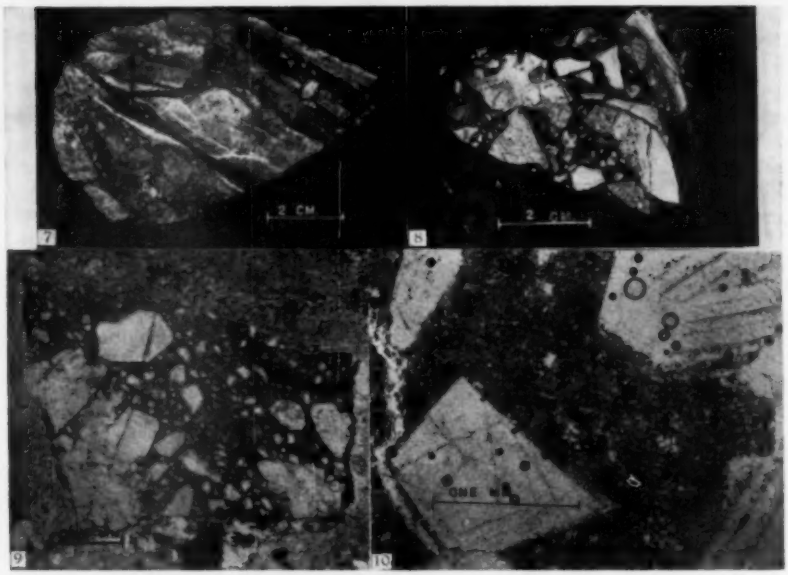


FIG. 7. Fluorspar breccia. Fragments of wallrock and fluorite are cemented by fine-grained fluorite, Mirabal Mine, Zuni Mountains, New Mexico.

FIG. 8. Fluorspar breccia showing two stages of brecciation, Mirabal Mine, Zuni Mountains, New Mexico.

FIG. 9. Fluorite breccia showing two stages of brecciation, Penber Shaft, Northgate, Colorado.

FIG. 10. Subhedral fluorite crystals are replacing a fine-grained quartz-hematite-fluorite mass. Ghost crystal boundaries are apparent. Red Dog Mine, Sandia Mountains, New Mexico.

Reticulated boxwork texture occurs in many Cordilleran fluorspar deposits. This texture may be considered as an intermediate form between undisturbed fluorite and collapse breccia. Although localized fluorite boxwork texture may be due to the action of descending surface waters, as described at Broken Hill, New South Wales, by Garrety and Blanchard (13), it is certain that many extensive boxwork zones have resulted from a stage of corrosion during primary mineralization. The effects of corrosion on fluorite and associated minerals are apparent in a majority of large fluorspar deposits. The resulting

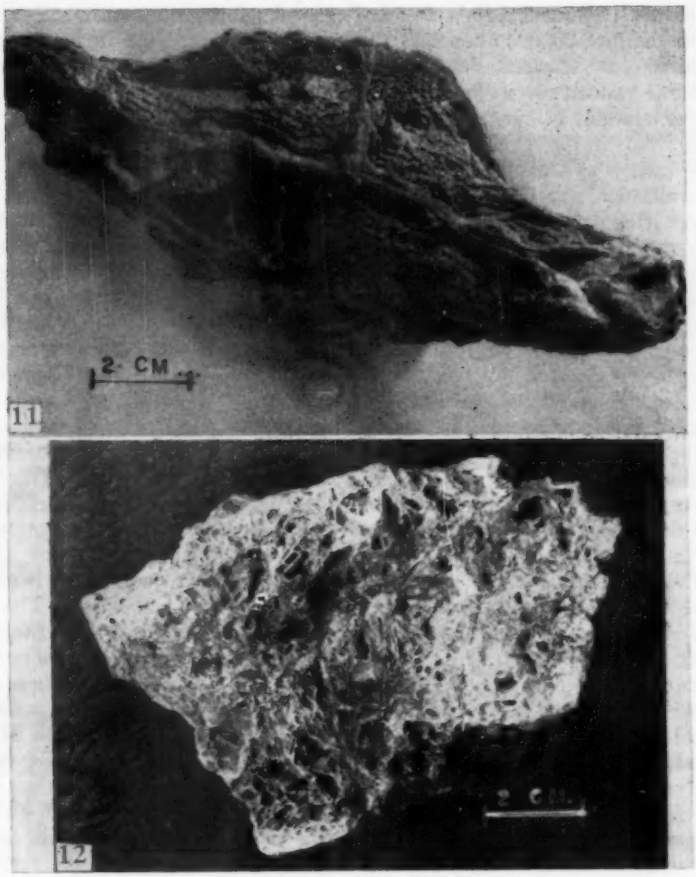


FIG. 11. Boxwork fluorspar, Thomas Range, Utah.
FIG. 12. Boxwork fluorspar, Northgate, Colorado.

boxwork, if further weakened, would be expected to collapse and form a breccia.

Near Challis, Idaho, at Keystone Mountain, a type of boxwork consists of thin partitions of limonitic jasper encrusted by coarse crystalline fluorite. In some areas the fluorite is in turn encrusted by quartz crystals. Open voids range from about an inch to nearly a foot in width. Locally, near the boxwork zones, the dolomite host rock contains a network of quartz and jasper veinlets. It is suggested that after silica filled reticulating fractures in the host rock, the dolomite was dissolved and fluorite was deposited in the resulting open boxwork.

In the Thomas Range, Utah, fine-grained fluorite commonly occurs in a coarse textured complex boxwork (Fig. 11). The partitions are composed of dense, slightly siliceous, fluorite. Openings in the boxwork are partly filled with fine-grained porous fluorite and very fine-grained "soapy" fluorite. Several generations of mineralization are clearly indicated, with each generation filling voids in the residual network remaining from the previous generation. Tiny quartz crystals were deposited on fluorite during the final stage of mineralization.

An irregularly developed boxwork (Fig. 12) is of widespread occurrence in fluorspar bodies at Northgate, Colorado. Massive fluorite zones and breccia zones are also found in the district. The boxwork partitions, composed of fine-grained crystalline fluorite, are encrusted by small fluorite cubes of a later generation. Nearly all of the voids are angular. A few of the voids have shapes that suggest former occupancy by a rhombohedral carbonate or by a feldspar; the remainder of the voids are irregular in outline and may have been occupied by fragments of the granitic wallrock. Two stages of fluorite deposition with intervening corrosion are indicated. Microscopic examination of a specimen from the Burro Chief Mine, near Tyrone, New Mexico, shows an antecedent network of fine-grained quartz in fluorite (Fig. 13). The quartz is partly embayed and replaced by fluorite. A change in equilibrium conditions could cause corrosion of fluorite and would result in a fine-textured quartz-fluorite boxwork.

Evidence supporting the thesis of corrosion during mineralization is noted at the Royal Flush Mine in the Hansonburg district, New Mexico. The deposit consists of a coarsely crystalline assemblage of barite, quartz, fluorite, and sulfide minerals in silicified limestone. Many of the large fluorite cubes that line cavernous openings in the orebody display etched surfaces. Corners and edges of many of the cubes have been corroded to the extent that the remaining crystals are nearly spherical. In one specimen from the Royal Flush Mine, corroded and rounded fluorite has been overgrown by a later galena crystal, thus proving that corrosion took place before primary mineralization had ceased. A few crystals display an interior zoning in which a core of etched and rounded fluorite is encrusted by a coating of quartz and barite. The latter is in turn covered by a clear fluorite cube with an etched and rounded surface.

Corrosion during mineralization has been cited by Locke, McKinstry, and many others. Similar intervals of corrosion have been postulated by Jahns for certain pegmatites and by Lacy and Hosmer (24) for sulfide deposits in Central Peru. Kerr (personal communication) has suggested that breccia zones in the Temple Mountain, Utah, uranium district are collapse features which have resulted from hydrothermal activity during an early stage of alteration and mineralization.

In the Dante Mine, at Cripple Creek, Colorado (Loughlin and Koschman, 28), a large pipe of collapse breccia contains quartz, fluorite, pyrite, and gold tellurides. Several stages of brecciation alternating with mineralization are demonstrated. The collapse is postulated to have resulted from corrosion during mineralization.

In the Cave-in-Rock district, Illinois, Grogan (21) has suggested that collapse structures in the rocks overlying fluorspar deposits are due to decrease in volume resulting from replacement of calcium carbonate by calcium fluoride. Gillerman (15, p. 60) has also suggested a volume decrease during mineralization in the Eagle Mountains, Texas. Volume change during replacement may be a contributing factor in the development of collapse breccias in Cordilleran fluorspar deposits. The major factor, however, is corrosion of hydrothermally introduced material and wallrock during mineralization.

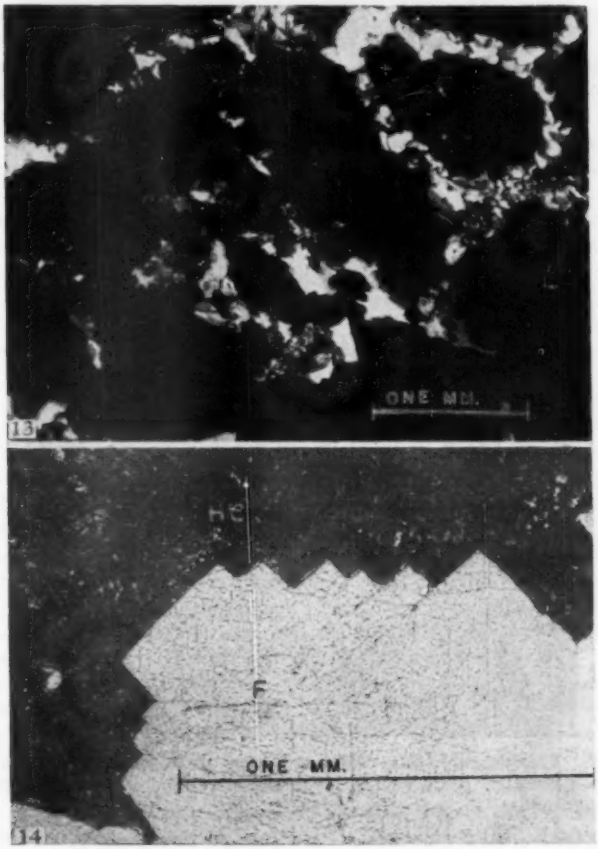


FIG. 13. Antecedent network of quartz (white and gray) in fluorite (black). Burro Chief Mine, near Tyrone, New Mexico. x-nicols. Burro Chief Mine, near Tyrone, New Mexico.

FIG. 14. Fluorite (F) replacing fine-grained hematite and calcite (HC). Zuni Mountains, New Mexico. Nicols at 45 degrees.

Breccia zones, breccia pipes, and chimney-like breccia shoots of various origins are common in Canada (50), Newfoundland (46), and Mexico (30), as well as in the western United States. Although many tabular breccia zones are of tectonic origin and one of the most effective structural controls for min-

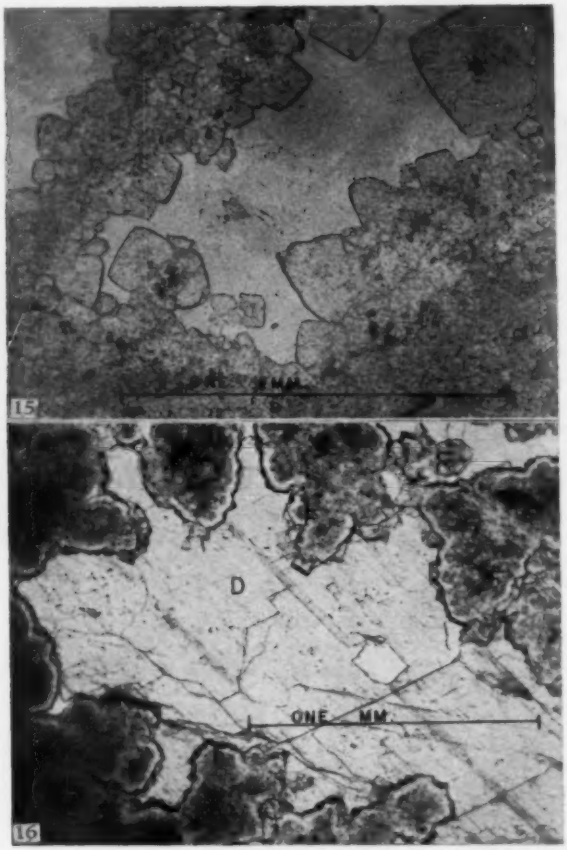


FIG. 15. Fluorite euhedrons replacing quartz, Zuni Mountains, New Mexico.
FIG. 16. Fluorite (F) replacing dolomite (D), Thomas Range, Utah. Photomicrograph.

eral localization is the presence of a tectonic breccia, many fluorspar breccias did not require continuous tectonic adjustment during mineralization. In pipe and chimney-like bodies, the breccia is irregular in distribution and appears to have resulted from collapse during mineralization. Collapse would be a natural consequence following corrosion of fluorite and the accompanying minerals.

Paragenesis.—Crustification and cockade structure, common features in fluorspar bodies as well as in many ore bodies that have formed in open fractures, provide criteria from which to determine the sequence of deposition.

In addition to its open-space characteristics in crustified veins, fluorite is an active mineral in replacing wallrock and gangue minerals, as shown in Figures 14, 15, and 16; thus, the earliest minerals in the sequence of crystallization may be nearly absent in the final fluorspar body.

Inclusions of foreign material in fluorite, generally in diameter ranging from sub-microscopic to a few tenths of a millimeter, are important factors in the paragenetic sequence. They may have been formed at various times during the sequence of crystallization. Stegmüller (41, p. 182) has suggested that inclusions of solid, liquid, or gaseous matter which form simultaneously with fluorite generally occupy positions parallel or at right angles to cubic or octahedral crystal faces, whereas material that was introduced at a later time tends to form a dendritic pattern along octahedral cleavages. Zones of solid inclusions arranged parallel to cube faces are abundant in dark purple fluorite crystals from the Todilto limestone near Grants, New Mexico; the included uraninite must have been formed contemporaneously with the fluorite, either as uraninite or, as suggested by Goldschmidt (19, p. 570), as uranium tetrafluoride. In most of the Cordilleran fluorite studied in thin section, inclusions that are large enough to be identified are randomly distributed and are composed of quartz, clay or micaceous minerals with embayed and corroded boundaries; they are remnants of earlier stages of mineralization or of wallrock.

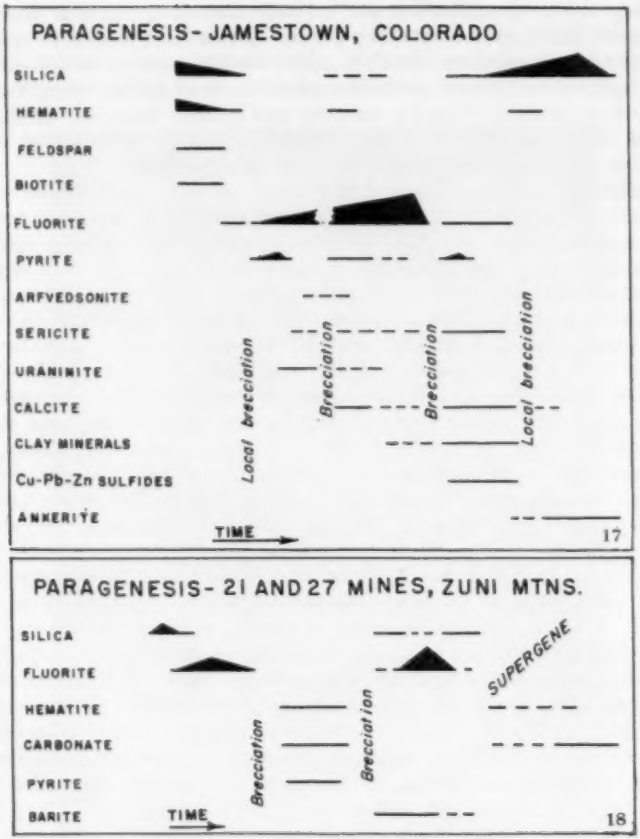
The paragenesis of Cordilleran fluorspar deposits differs from district to district. In most instances there are repetitions within the sequence of mineral deposition which are associated with stages of brecciation. In keeping with inferences from textural and morphologic studies of fluorspar deposits, there is an indication of disturbance during mineralization. Paragenetic diagrams shown in Figures 17, 18, 19, 20 and 21 illustrate typical sequences.

Insofar as generalization regarding paragenesis is possible, it appears that the sulfide minerals are dominantly associated with an early stage of mineralization; barite is characteristic of the closing stages. Quartz has a long history in most deposits; it may precede, accompany, or follow fluorite. Carbonates are dominantly late minerals, but because they are easily replaced, they may be of greater importance in earlier stages than is apparent.

Zoning.—Fluorite may be deposited through a large range of temperature, pressure, and concentration; it does not indicate a specific environment of deposition unless accompanied by other minerals.

Barite, one of the most common associates of fluorite in the Western United States, has a zonal relation in many districts. Barite generally lies outside of the fluorite zone, specific examples being in the Zuni Mountains, New Mexico, in the Organ Mountains, New Mexico (12, p. 138), and in the Sandia Mountains, New Mexico. At Capulin Peak, in the Sandia Mountains, a fluorspar vein changes to a barite vein as it leaves the area of most intense mineralization.

Wallrock Alteration.—Silicification, the most common type of wallrock alteration associated with fluorspar, is illustrated by the Keystone Mountain and



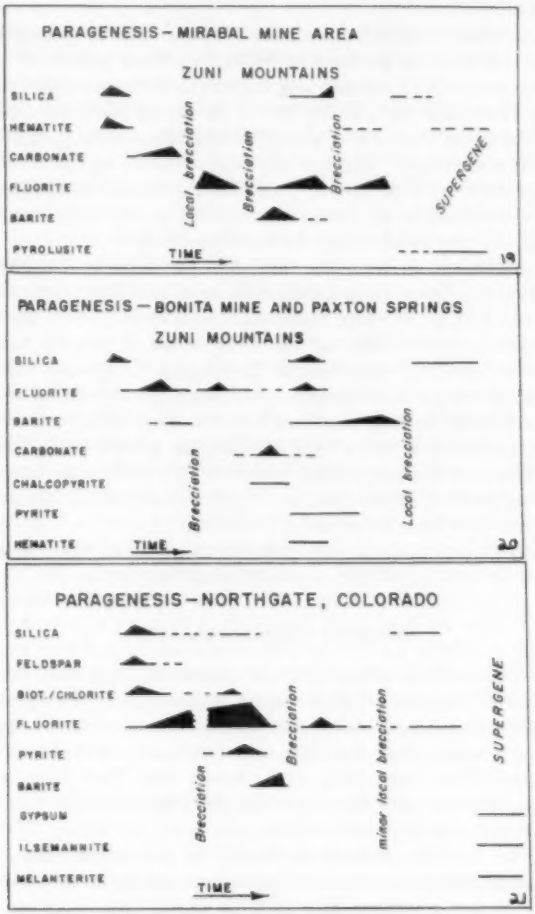
FIGS. 17 AND 18.

Meyer's Cove districts, Idaho, where fluorspar bodies are found within silicified zones. The relationship is not without exception, however, in that a few silicified zones in each of the above mentioned districts do not contain fluorspar bodies.

"Jasperoid" type silicification (39, p. 219) is widespread in the limestones around fluorite-base metal bodies in the Hansonburg district, New Mexico. Successive stages in silicification are apparent, beginning with an outer envelope of disseminated quartz needles, continuing through an intermediate zone of intergrown quartz and calcite, and culminating in an inner zone of silica rock with crystals of fluorite, barite, and galena.

In most of the fluorspar district studied by the writer, there is some silicification in the vicinity of the mineralized bodies. Silica in altered zones and in

ore bodies may have been derived from wallrock in the vicinity as suggested by Schmitt (36) who points out that ore-forming solutions are capable of causing desilication of alumino-silicate walls during an early stage of wallrock alteration. A change from desilication to silicification during hydrothermal activity would result from saturation of silica in the solutions. Causes of saturation which have been suggested are decrease in temperature, decrease in pressure, and change in pH. It has been shown by several writers, that although silica may be dissolved by all natural waters, it is more soluble in alkaline than in acid waters. The suggested derivation of silica from wallrock



FIGS. 19, 20, AND 21.

during early stages of hydrothermal activity is in agreement with the observation that quartz is deposited throughout the usual mineral sequence in fluorspar deposits and thus does not depend entirely upon the same source conditions as the fluorite.

Inasmuch as most Western fluorspar deposits are associated with granitic or alkalic rocks which are high in hyperfusible components, the effects of deuteric alteration are widespread and are difficult to distinguish from those of alteration during the hydrothermal stage. Sericitization, as an example, is noted in almost all wallrock of fluorspar deposits, but it does not seem to form a recognizable pattern that can be identified with the epoch of fluorspar mineralization.

Carbonatization, although associated with the interior portions of fluorspar bodies, does not seem to produce pervasive wallrock alteration. Dolomitization of limestone is not an outstanding feature in fluorspar districts. The dolomites of the Thomas Range, Utah, and of Keystone Mountain, Idaho, seem to predate the fluorspar bodies and are apparently associated with regional rather than local dolomitization. Some of the coarse calcite in the Todilto limestone of the Grants district, New Mexico, may be a genetic associate of fluorite, but the irregular distribution of coarse calcite over a very large area is explained more logically by recrystallization during diagenesis.

Among references to wallrock alteration in Western fluorite-bearing deposits, mention has been made of chloritization and propyllitization at Castle Dome, Arizona (33, p. 91) and Jamestown, Colorado (18, p. 20–25).

Argillic alteration of feldspars has been noted in a great many fluorspar districts and is especially apparent at Northgate, Colorado. It is not, however, as widespread as silicification. Although an alunitic zone lies in close proximity to fluorite-bearing uranium deposits near Marysvale, Utah, alunitization is not generally a major type of wallrock alteration in fluorspar areas.

Feldspathization has been noted in the vicinity of fluorite-bearing bodies in the Blue Wing district, Idaho (1), at Cripple Creek, Colorado, and in various localities in the San Juan Mountains, Colorado.

Greisen zones and disseminated fluorite in granitic rocks do not generally bear any distinct spatial relation to commercial fluorite bodies.

ORIGIN OF FLUORSPAR DEPOSITS

Regional Control.—A comprehensive statement regarding the distribution of fluorite in the Western United States was made by Lindgren (26) who noted that fluorite is scarce along the Pacific Coast, relatively uncommon in Idaho, Utah, Nevada, and Arizona, and most abundant along the Eastern Rocky Mountain front, especially in Colorado and New Mexico. Lindgren proposed that there is a genetic connection with hyperfusible-rich alkalic rocks in the Eastern portion of the Cordillera.

A genetically similar pattern is shown on a smaller scale in the Front Range, Colorado; alkalic Laramide intrusives are associated with fluorspar. Lovering and Goddard (29a) show that Laramide intrusives on the Western slope of the Front Range have a characteristic composition between that of

augite diorite and quartz monzonite, whereas on the Eastern slope alkalic syenite, alkalic trachite, and bostonite are more common. Fluorite is more abundant on the Eastern slope and is of commercial importance at Jamestown in the vicinity of sodic granite and bostonite intrusives. Sodic granite at Jamestown and bostonite at Central City carry fluorite as an accessory mineral.

Lindgren explained the preponderance of fluorite and alkalic rocks along the Eastern margin of the Cordillera in terms of regional magmatic differentiation of a large source at depth, resulting in earlier intermediate magmas along the Western margin and later alkalic and felsic magmas along the Eastern margin. The explanation is in keeping with the concept that orogeny progressed from West to East through the Cordilleran system, although the regional magma invoked by Lindgren does not appear probable. New fluorspar discoveries have served to corroborate Lindgren's proposals as regards regional distribution.

The fluorite-bearing zone of the Eastern margin of the Cordilleran area has the characteristic of being integrated with local ore types in various districts. In southwestern New Mexico, fluorite accompanies copper mineralization; fluorite accompanies lead-zinc mineralization in the Rio Grande Valley area, New Mexico. Fluorite is associated with gold and sulfides in the San Juan Mountains, Colorado, with tellurides at Cripple Creek, Colorado, and with gold and sulfides at Jamestown, Colorado. Farther to the west, fluorite does not accompany the copper ores of Bingham Canyon, Utah, or Ely, Nevada; it is not widespread in the gold and basemetal ores of the Pacific Coast and Basin-Range Regions. It must be concluded that fluorite is more characteristic of a region than of a particular genetic family of ore deposits.

The Western Cordilleran region may be divided into several tectonic units; two units are especially significant to the distribution of fluorite. A Western unit, including California, Oregon, Washington, Nevada, Idaho, western Utah, and Arizona is characterized by thick sediments, an intense compressional tectonic pattern, and a dearth of exposed Precambrian basement uplifts. This unit, constituting the area of the Paleozoic Cordilleran geosyncline, is relatively poor in fluorite. An Eastern unit in Montana, Wyoming, Colorado, and New Mexico is characterized by a thin sedimentary cover, a combined compressional and vertical tectonic pattern, and widespread basement uplift areas. This Eastern unit, a shelf area during Paleozoic time, is relatively rich in fluorite. It would seem that the tectonic and sedimentary patterns, as well as the igneous pattern, have significance in the distribution of fluorite. Inasmuch as many of the fluorspar deposits are in Precambrian rock, it is further possible that the pattern has been inherited from Precambrian time, with the present deposits a Laramide-Tertiary expression of a long-existing regional concentration.

An apparent tendency for fluorine to remain in a specific region during several geologic cycles has been discussed in the section on geochemical considerations. It is noteworthy in this connection that Wilson (49), and others have suggested that the sialic continental areas were developed from a simatic crust by repeated differentiation through sedimentation, metamorphism, and fusion. By Wilson's concept, the North American continent has grown ocean-

ward from several nuclei by accretion of successively differentiated material. If the development of the Western Cordilleran region has involved growth by accretion, the Precambrian basement of the Rocky Mountain front has undergone a longer history of differentiation than have the basement rocks of the West Coast. Considering that fluorine is difficult to dislodge from a region, it is possible that a Precambrian belt of fluorine-bearing rocks in the area of the Rocky Mountains would influence later cycles of differentiation and would be maintained through successive generations of igneous, sedimentary, and metamorphic rocks.

The regional pattern of fluorite in the western United States, considered from the standpoint of continental growth by accretion, suggests that a long-standing fluorine-rich area exists in the basement rocks and has been incorporated into the Mesozoic and Tertiary mineralization pattern. The distribution of the more highly differentiated alkalic rocks is in close correlation with the distribution of fluorite because the eastern margin of the Cordilleran region has had a longer history of differentiation. The limits of the fluorine-rich basement area are not well defined, inasmuch as the basement rocks are deeply buried in most of the western part of the Rocky Mountain region. Possibly several zones of relatively high fluorine content exist in the basement complex and some were affected by Tertiary mineralization, including one zone in Colorado and New Mexico, one in Montana, and a smaller zone in southwest Utah.

The Nature of the Mineralizing Fluid.—A wide range in physical and mineralogical environment is evident in fluorite-bearing deposits; commercial deposits are, however, characterized by shallow depth of formation.

Paragenetic studies indicate that sulfide minerals, uraninite, and hematite are associated with early stages in fluorite deposition, that barite is characteristic of later stages, and that carbonate minerals and silica appear in all stages. Silica is also important in wallrock alteration adjacent to fluorspar bodies.

A great majority of commercial fluorspar bodies are classified with those hydrothermal deposits emplaced within a zone of rapid temperature and pressure changes deriving from direct surface influence. Schmitt (37) has suggested that magmatic products join meteoric waters to form shallow-seated ore deposits; a process of this type may be readily applied to the fluoritic uranium ores near Grants, New Mexico, to fluorite deposits at Thermal Springs in Colorado and New Mexico, and to a great many of the largest Cordilleran fluorspar deposits. A striking example of fluorite emplacement by combined surface and magmatic fluids is presented by Schneider (38), who has proposed that submarine volcanic emanations acted during precipitation of marine limestone to form an extensive belt of commercial fluorspar deposits in the Calcareous Alps of Bavaria and the Tyrol.

The common occurrence of fluorspar in breccia pipes and in brecciated pipe-like zones within vein systems points toward active solution and collapse during mineralization. Corrosion of early generation vein material, of wallrock, and of fluorite has produced a boxwork texture in many fluorspar deposits. It is suggested that boxwork texture represents an intermediate stage between the simple undisturbed texture of veins and the breccia texture of

pipes. The implication regarding the genesis of fluorspar bodies is that there are alternating times of stability and instability during which fluorite and certain of the accompanying minerals are first deposited and then taken into solution.

The instability that attends fluorite deposition results from a change in equilibrium between solutions and the deposited minerals. This may be due to changes in the solutions at the source, to reaction between the solutions and the walls of the conduit, or to temperature and pressure changes in portions of the system.

A type of instability that may apply to fluorspar deposits is discussed by Lacy and Hosmer (24, p. 78) in their description of collapse breccias and boxwork textures in sulfide bodies in Central Peru. They maintain that the deposition of carbonates, sulfides, and barite would change to dissolution following a drop in temperature during the later stages of mineralization. The action is suggested by a laboratory process in which decreasing of temperature causes a decrease in the concentration of the sulfate ion, an increase in the concentration of the sulfur ion, and an increase in the solubility of calcium carbonate.

Lovering and others (29, p. 56) have described an instance in the East Tintic district, Utah, in which barite is found in several generations with each generation showing evidence of corrosion before deposition of associated quartz. As in many fluorspar deposits, a delicate balance between corrosion and deposition is indicated. Lovering proposes that rapid fluctuations in pressure such as would accompany intramineralization faulting would affect the amount of free carbon dioxide in solution, and thus cause either precipitation of barite under a pressure decrease or corrosion of barite under a pressure increase.

Goddard's (18, p. 25) explanation of collapse in the fluorspar bodies at Jamestown, Colorado, holds that a period of fluorite deposition was followed by a change in the character of the solutions, possibly from alkaline to acid, and consequent corrosion of material already deposited.

Regarding stages of corrosion and instability in the genesis of fluorspar bodies, it is apparent that changes in composition and pH of the solutions or changes in temperature and pressure are likely causes. A change in physical state from liquid to gas would be effective, but would not be necessary in order to accomplish corrosion.

Lowering of temperature in a fluorite-depositing system due to contact with surface waters or by diversion of fluids from a partly filled conduit to a new one would cause an increase in the solubility of carbon dioxide and effect a corrosion of previously deposited calcite. Continued corrosion of calcite would lead to collapse and to an attendant radical departure from equilibrium for the entire system. Even without immediate collapse, the addition or removal of carbon dioxide would be expected to disturb the mass action relationships between the fluid and the minerals already deposited.

Causes for instability outside of the immediate depositional environment would involve changes in temperature, pressure, and concentration in the

source or conduit area. A hydrothermal system would be expected to begin with a stage of heat transfer to the walls, culminate in a high temperature stage in which the walls of the system are chemically and physically in equilibrium with the solutions, and then wane in a cooling stage. The opportunities for temporary instability during the cycle are numerous and do not require a profound disturbance to "trigger" the action.

The processes considered would be applicable to nearly all types of ore deposition in the near surface zone. Its apparent prevalence in fluorspar deposits is probably due to the corrosive nature of fluorine and hydrofluoric acid in comparison to the more common sulfuric and carbonic acids.

Studies in the transportation of uranium and fluorine have been made by Green and Kerr (20, p. 67) relative to uraninite deposits at Marysvale, Utah. They state that uranium probably entered the vein system as UF_4. Upon contact with water in its ascent, the vapor pressure decreased and the fluid hydrolyzed to form UO_2 and HF. The released HF then reacted with calcium compounds in the wallrock to form fluorite. The paragenetic sequence of commercial fluorspar deposits indicates that the deposition of sulfide and oxide minerals is related to the deposition of fluorite. The reaction cited by Green and Kerr for deposition of uraninite may be of significance in this respect. Deposition of metallic compounds from solution would liberate HF to form fluorite in the immediate vicinity or slightly beyond the metallic zone. The occurrence of barite in the later stages of mineralization would indicate that sulfate tends to persists in the cooling solutions after removal of fluorine.

It has been considered by Bowen (6) and others that silica may be carried in hydrothermal solutions as SiF_4 and that addition of water or lowering of pressure will cause the formation of silica with following equilibrium relation: $Si_4 + 2H_2O \rightleftarrows Si_2 + 4HF$. A certain amount of early vein quartz and early wallrock silicification may have formed in this manner. It is, however, noteworthy that silicification is not confined to the stage of fluorite deposition. It is, in fact, practically ubiquitous in the mineralization sequence, and must therefore have been transported in various ways.

Summary.—In the Cordilleran region of the United States, the genetic features of fluorspar deposits are integrated with those of Mesozoic and Cenozoic ore deposits. Fluorspar deposits are greatly concentrated along the Eastern margin of the Cordillera in a pattern that may have existed since Precambrian time, the present deposits being in large part derived from deeply buried basement rocks. Fluorite is a relatively common minor gangue mineral in nearly all types of ore deposits, but it is of commercial importance mainly in shallow-seated deposits.

Major fluorspar deposits are commonly characterized by corrosion and collapse features that reflect recurrent instability in the fluorite-depositing fluids. It is believed that instability results from minor disequilibria either at the source of the fluids or within the depositional system. Similar occurrences are noted in many non-fluoritic ore deposits, but the corrosive processes are not generally so intense as those which involve fluorine.

Fluoritic solutions are similar to other types of ore-bearing hydrothermal

solutions in composition and derivation. The effects of meteoric water, and complex near-surface temperature-pressure variations are especially important.

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Nov. 18, 1957

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TEMPERATURES AND DEPTH OF FORMATION OF SULFIDE ORE DEPOSITS AT GILMAN, COLORADO[1]

TOM G. LOVERING

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ABSTRACT

The temperature of formation of sulfide replacement deposits in the Leadville limestone at Gilman, Colorado, has been estimated from mineralogical studies, iron content of sphalerite, pyrite geothermometry; fluid inclusion studies, and thermoluminescence glow curves.

Deposition of the sulfide ores was preceded by hydrothermal dolomitization of the carbonate country rocks. This dolomitization seems to have taken place at temperatures that attained a maximum of slightly more than 300° C and then declined to about 230° C before dolomitization ceased. The texture and composition of the early sphalerite ore indicates that it was emplaced at temperatures between 500° and 600° C. Later ore minerals were deposited at successively lower temperatures, and hypogene ore mineralization probably closed with temperatures below 150° C.

It is estimated that the depth of formation of these deposits probably was not less than one mile nor more than three miles. These calculations are based on a reconstruction of the stratigraphic column at the time of mineralization, extrapolation of the geothermal gradient, comparison of fluid inclusion temperatures with glow curve temperatures, and estimated crushing strength of rocks surrounding open pre-ore cavities.

INTRODUCTION

At Gilman, in central Colorado, the Leadville limestone of Mississippian age has been extensively replaced by sulfide deposits consisting largely of pyrite, sphalerite, chalcopyrite, and galena. Samples of these ores, associated gangue minerals, and country rock have been investigated in an attempt to

[1] Publication authorized by the Director, U. S. Geological Survey.

establish the temperature range through which mineralization took place, and together with other lines of evidence, to estimate the probable depth at which it occurred.

The first comprehensive geologic study of this area was made by Crawford and Gibson (4) in the early 1920's. Later the geology and ore deposits of the Gilman district were briefly summarized by Lovering and Behre (15) for the Sixteenth International Geological Congress. In 1939 T. S. Lovering and Ogden Tweto began a thorough study of the geology and ore deposits of the Minturn quadrangle, with particular emphasis on the Gilman district. The field work for this study was largely completed in 1942, and a preliminary report was placed on open file by the U. S. Geological Survey in 1944 (16). A summary of the geology and ore deposits at Gilman, based on this work, was later published by Tweto and Lovering (28). In 1955 the author began a laboratory study of the rock and ore samples collected by Lovering and Tweto. This paper is an outgrowth of that study, but also includes much information taken from the open file report.

Gilman is at the base of the northeastern flank of the Sawatch Range, perched on a narrow shelf 1,000 feet above the upper canyon of the Eagle River. It is about 75 miles airline west of Denver and about 30 miles north of Leadville (Fig. 1).

Since its discovery in 1879 the Gilman district has developed into one of the major producers of copper and zinc in Colorado, and has also had a substantial output of gold, silver, and lead. The total value of the ore produced from this district is more than 180 million dollars.

The district is on the southwestern limb of a northwest plunging regional syncline, filled with Paleozoic sedimentary rocks, that separates the exposures of Precambrian crystalline and metamorphic rocks in the Sawatch Range from those in the Gore Range 7 miles to the northeast. The long gently dipping southwestern limb of the syncline gives way abruptly to a steep northeastern limb that is truncated by the Gore Fault, which has been traced for more than 30 miles along the western flank of the Gore Range.

The ore bodies at Gilman occur within an area of about three square miles. The major production has come from deposits in the Leadville limestone of Mississippian age and in the underlying Dyer dolomite member of the Chaffee formation of Late Devonian age. These formations are separated from the Precambrian basement rocks by about 250 feet of quartzite, shale, and dolomite. A quartz latite porphyry sill, 30–60 feet thick, which has been highly altered, was intruded near the base of the Pennsylvanian rocks overlying the Leadville limestone. This sill, which is probably of early Tertiary age, is older than the ore deposits.

The ore deposits are of three main types: vein deposits in the Precambrian rocks, bedding replacement or manto bodies in the quartzites and dolomites, and pipes or chimney deposits in the dolomites. Most of the ore in the district was taken from manto deposits in the upper part of the dolomitized Leadville limestone, beneath the relatively impervious cap rock formed by the quartz latite porphyry and the overlying Pennsylvanian black shale. This report is confined to a study of the ore bodies in the carbonate rocks, and of the

hydrothermally altered Leadville limestone and Dyer dolomite member of the Chaffee formation in the Gilman district.

The ore mantos in the Leadville limestone are long, sinuous, and rudely elliptical in cross section; they are as much as 300 feet wide, 150 feet thick,

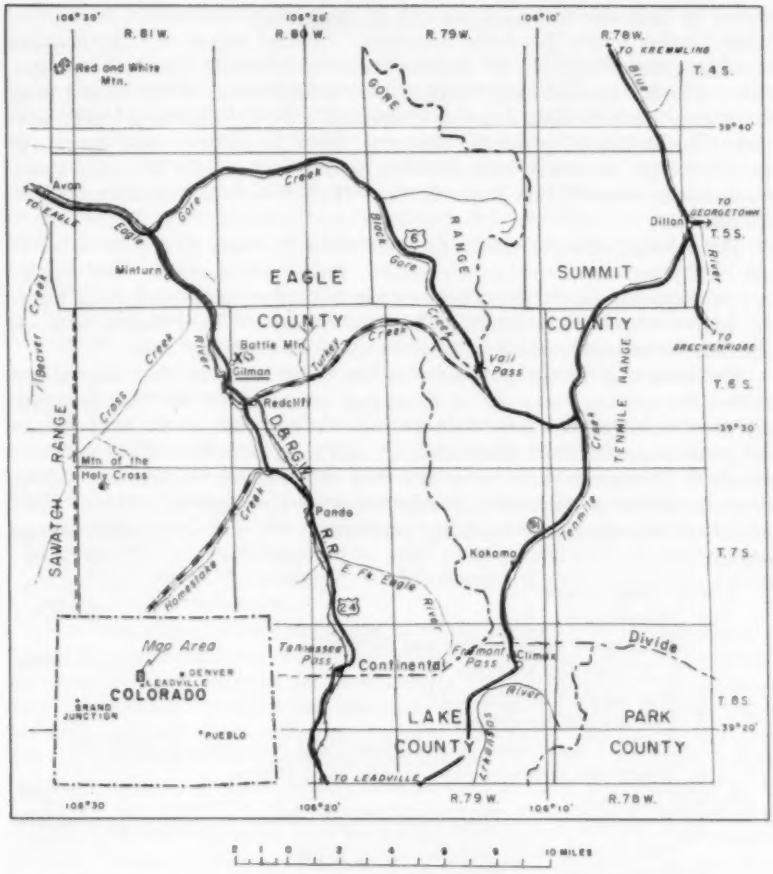

FIG. 1. Index map showing location of the Gilman district, Colorado.

and several hundred feet long. These mantos were localized by previously formed solution channels. The dominant minerals are sphalerite, pyrite, and manganosiderite; minor minerals include galena, chalcopyrite, barite, dolomite, and quartz. The chimney ore bodies are thickened roots of the mantos, and

project downward from them into the underlying Dyer. These chimney deposits consist of barren pyrite with irregular bodies of pyritic ore in which chalcopyrite, copper-silver sulfosalts, and argentiferous galena are present in widely differing amounts. The chimneys are the major source of copper and silver in the district.

Prior to the introduction of sulfides, the rocks in this area were extensively altered by hydrothermal solutions rich in magnesium that caused widespread dolomitization of the Leadville limestone. Several stages of dolomitization have been recognized. In the earliest stage the Leadville limestone was converted into fine-grained dark dolomite over a large area; in the second stage alternating bands of dark dense dolomite and white crystalline dolomite were formed in the vicinity of the ore deposits (Zebra dolomite of local usage); in the third stage, a pinkish buff dolomite breccia was formed in a thin stratigraphic zone about 95 feet above the base of the Leadville limestone near the ore deposits.

The dolomitized rock was later leached along fractures by pre-ore solutions and locally converted to dolomite sand. Some open channels were formed and subsequently partly filled with rubble from their sides and roofs before the advent of sulfide mineralization, which seems to have begun with the deposition of manganosiderite and pyrite along solution channels.

Speculation as to the chemistry of the early solutions that altered and leached the country rocks and of those that deposited the later barren pyrite and ore should be guided by the best available estimates of the temperatures and pressure under which they acted, as well as by the chemical changes they produced. Fortunately the rocks and ores of the Gilman district are susceptible to a variety of methods of geothermometric investigation, some of which can also yield estimates of confining pressures. The following methods were employed:

I. *For temperature*
   1. Mineralogy
      a. Mineral assemblages and sequence
      b. Exsolution textures
   2. Iron content of sphalerite
   3. Pyrite geothermometer
   4. Filling temperature of two-phase fluid inclusions in dolomite.
   5. Thermoluminescent glow curves of dolomite

II. *For pressure*
   1. Thickness of overlying rock at time of ore deposition as estimated from stratigraphy and paleogeography.
   2. Extrapolation of the geothermal gradient
   3. Pressure correction needed to reconcile fluid inclusion filling temperatures with temperature estimates from glow curves
   4. Estimated crushing strength of dolomite

The results of these various investigations suggest that much of the early dolomitization took place at moderate temperatures, in the neighborhood of 250° C; that the first ore-stage solutions were much hotter (between 500°

and 600° C) at the beginning of mineralization, and that as mineralization progressed the solutions gradually cooled to below 150 degrees. The pressure during mineralization was probably not less than 400 nor more than 1,200 atmospheres, corresponding to a lithostatic pressure of between one and three miles of overlying rock.

GEOTHERMOMETRY METHODS

A great variety of methods for estimating the formation temperature of rocks and minerals has been proposed during the past half century (8). Unfortunately, no single method has been found that is applicable to the common ore and gangue minerals of sulfide ore deposits, and that is reasonably precise, widely applicable, and whose underlying assumptions are universally accepted. Nevertheless, where several techniques give concordant results, as in the present investigation, the approximate temperatures can be estimated with a considerable degree of confidence. Furthermore, the reliability of the various methods can be evaluated.

Mineral Assemblage.—Certain ore and gangue minerals are generally regarded as diagnostic of high temperatures and others of low temperatures (13). A large group of minerals that can form through a wide range of temperature are considered to be "intermediate" when not associated with those indicative of high or low temperature.

The primary ore bodies of the Gilman district are composed of minerals belonging to the intermediate group. Unfortunately there is no general agreement as to where the temperature boundaries for this group should be placed. The general sequence of formation observed at Gilman, from high temperature to low, accords well with that given by Lindgren (12) and also by Park (18). The sequence in the Gilman ores, somewhat generalized, is as follows: 1. siderite, 2. pyrite, 3. sphalerite, 4. chalcopyrite, 5. tetrahedrite, 6. galena, 7. gold and silver tellurides.

The foregoing mineral assemblage, with the exception of the tellurides, is typical of mesothermal ore deposits as defined by Lindgren (12). The formation temperature of some of the early Gilman ore minerals however, seems to have substantially exceeded the arbitrary upper limit of 300 degrees, which has been assigned to mesothermal deposits (12, 13). The assumption that this sequence represents deposition from successively cooler solutions will be supported in part by supplementary data on some of these ore minerals.

There is a possibility that the pyritic copper-silver ores of the chimney deposits represent a separate later period of mineralization from the zinc manto deposits, but even if they do, the concept of the general sequence, representing deposition from successively cooler solutions, still appears to be valid.

The gangue minerals of the Gilman ore bodies may also furnish evidence as to the temperature attained by the hydrothermal solutions. If siliceous limestones or dolomites are strongly heated they become metamorphosed with the formation of lime-silicate minerals in a progressive sequence with increasing temperature (2). Harker and Tuttle (7) show that quartz of siliceous limestone or dolomite will react with calcite to produce wollastonite if the

temperature exceeds 600° C at confining pressure of about 4,000 pounds per square inch, which is probably a greater pressure than that under which the Gilman ore deposits formed. Theoretically, tremolite, fosterite, and diopside should form at even lower temperatures and pressures (2). The absence of these products of thermal metamorphism in siliceous carbonate rocks surrounding the ore bodies at Gilman suggests that mineralization temperatures did not exceed 600° C for any appreciable length of time, if at all.

Mineral Textures.—Among the many mineral pairs for which exsolution textures can be used to establish temperature limits, the Gilman district has sphalerite-chalcopyrite, bornite-tetrahedrite, and pyrrhotite-chalcopyrite.

Much of the Gilman sphalerite that is high in iron contains tiny, randomly oriented, blebs of chalcopyrite forming a texture similar to that studied by Buerger (3), who showed that prolonged heating of the sphalerite at temperatures above 400° C caused the chalcopyrite to disappear, going into solid solution in the sphalerite, and that on slow cooling the chalcopyrite blebs reappeared at about 350° C. Samples of sphalerite from Gilman that are low in iron show no exsolved chalcopyrite, even though chalcopyrite is the mineral next in the paragenetic sequence after sphalerite in these samples. These facts suggest that deposition of much of the sphalerite took place above 350° C, but that some of the later sphalerite formed below that temperature.

Unfortunately, temperature estimates based on the exsolution of chalcopyrite from sphalerite can no longer be considered very reliable. Recent investigation of the Cu-Fe-Zn-S system by Kullerud indicates that there may be considerable variation in this exsolution temperature corresponding to variations in the copper content of the sphalerite (P. B. Barton, written communication). Nevertheless, temperature estimates on sphalerite samples from the Gilman district, based on the presence or absence of exsolved chalcopyrite, are in fair agreement with the more specific minimum estimates from the iron content of the same sphalerite samples. This method is discussed in the following section.

Sphalerite in the copper ores at Gilman is commonly followed by chalcopyrite, which is partially replaced by later tetrahedrite. Some specimens of such ore show alternating rhythmic bands of tetrahedrite and chalcopyrite, and similar alternating bands of tetrahedrite and bornite. The observed textures are strongly suggestive of simultaneous deposition, yet none of these minerals show exsolution textures with respect to the others. Edwards (5) established the temperature at which tetrahedrite exsolves from bornite at about 275° C. It thus seems safe to assume that the minerals in these samples were deposited below that temperature.

Some of the copper ore samples from the chimney deposits contain penecontemporaneous pyrrhotite and chalcopyrite, yet the chalcopyrite contains no exsolved pyrrhotite. Experimental work by Borchert (1) indicated that, in a system containing excess iron, chalcopyrite formed above 250°C contains sufficient iron for pyrrhotite to exsolve when the chalcopyrite cools. The absence of exsolved pyrrhotite in the chalcopyrite of these samples may thus be taken as evidence that the chalcopyrite formed below 250° C.

Argentiferous galena of the chimney ores was deposited later than most, if not all, of the copper sulfides, and is the youngest of the primary ore minerals in most places. Polished sections of this galena characteristically contain numerous tiny blebs and amoeboid inclusions of hessite (Ag_2Te), but neither mineral cuts the other. This texture can be interpreted to indicate contemporaneous deposition of galena and hessite. Dana's system of mineralogy (17) states that the high temperature form of hessite is isometric and that on cooling it inverts to a monoclinic (?) form which is anisotropic; twinning in the low temperature form disappears when the mineral is heated to 149.5° C. This temperature probably varies somewhat, depending on the composition of the hessite. Ramdohr (19) observed mimetic twinning in anisotropic hessite and concluded that the mineral was formed at some temperature in excess of 150° C. The hessite in galena from Gilman is anisotropic with patchy polarization colors and rarely exhibits twinning. In one or two of the samples in which twinning is evident two sets of twin lamellae, intersecting at right angles, can be observed; these may represent mimetic twinning derived from high temperature isometric hessite. The fact that the vast majority of hessite inclusions in galena from Gilman do not show mimetic twinning suggests that most of this hessite formed below 150° C, although it may have exsolved from galena that was deposited at a higher temperature. Nevertheless, some of the hessite inclusions are cut by later primary native gold, thus indicating that ore deposition continued beyond the time at which the hessite formed.

For the minerals: sphalerite-early chalcopyrite, tetrahedrite-bornite, chalcopyrite-pyrrhotite, and hessite, we thus have the following estimates of formation temperature from a study of mineral textures.

sphalerite-early chalcopyrite	$> 350°$ C
tetrahedrite-bornite	$< 275°$ C
chalcopyrite-pyrrhotite	$< 250°$ C
hessite	$< 150°$ C

This evidence suggests that crystallization of most of the primary ore at Gilman probably began at temperatures above 350° and ended below 150° C.

Iron Content of Sphalerite.—An independent estimate of the temperature of formation of sphalerite can be made from its iron content. For many years geologists have believed that light-colored sphalerite with a low iron content was formed at low temperatures and that dark iron-rich sphalerite known as marmatite was formed at moderate to high temperatures. Recent experimental work by Kullerud (11) has put this qualitative concept on a quantitative basis. He established the iron content of sphalerite formed in equilibrium with pyrrhotite at temperatures from 400° C to 900° C and extrapolated the resulting curve down to 140° C (Fig. 2). Kullerud (11) points out that this method is applicable only where there is evidence that an excess of FeS was present in the solutions from which the sphalerite was deposited. Such evidence could be furnished by the presence of pyrrhotite formed contemporaneously with sphalerite.

Although pyrrhotite is rare in the Gilman ores, pyrite is closely associated with the high-iron sphalerite. Textural relations between the two minerals indicate that deposition of sphalerite began before deposition of pyrite ended.

Analyses of 3 sphalerites from the Gilman district for zinc and iron are given in Table 1. These sphalerites do not contain any pyrite that is visible

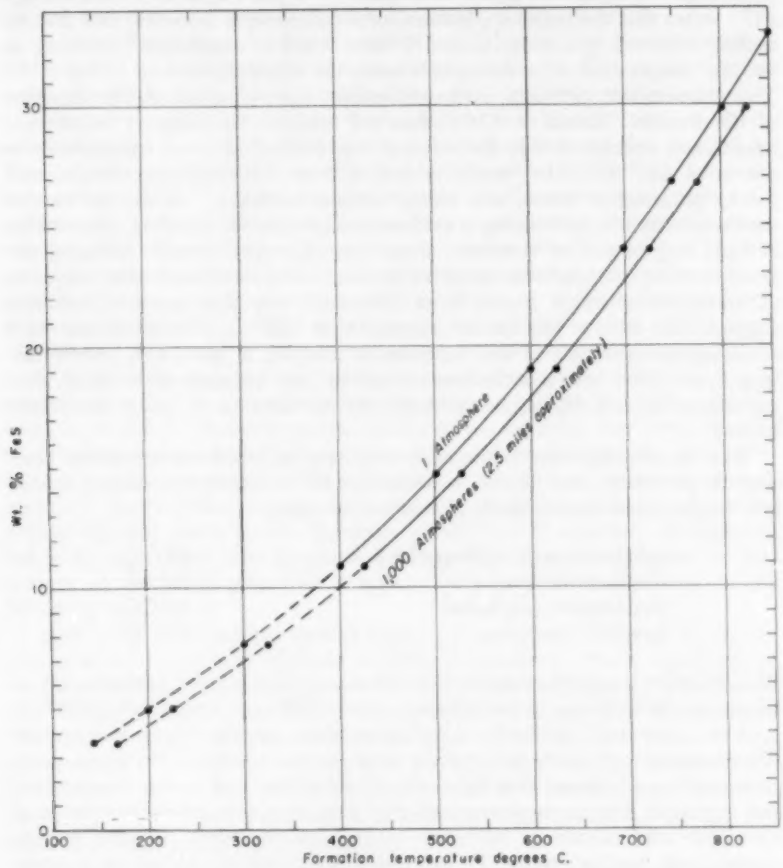

FIG. 2. Variation of the FeS content of sphalerite with temperature and pressure (Data from Kullerud (11)).

under 100× magnification, but samples 61T40 and 41T41 contain tiny blebs of chalcopyrite. If these analyses are converted into equivalent ZnS and FeS, the minimum formation temperatures of the samples can be estimated from Kullerud's graph, shown in Figure 2, by assuming that pressure at the time

TABLE 1

ZINC AND IRON CONTENT OF SPHALERITE SAMPLES FROM GILMAN[1]

No.[2]	Zn percent	Total Fe as percent Fe_2O_3
61 T 40	53.64	15.36
12 T 41	58.68	9.18
41 T 41	55.89	13.49

[1] Edward C. Mallory, analyst.
[2] Laboratory number and location given in appendix.

of formation did not exceed 1,000 atmospheres. The results of these calculations are shown in Table 2.

The temperatures shown in Table 2 represent minimum estimates because of the presence of pyrite rather than pyrrhotite associated with sphalerite. The absence of wollastonite in the siliceous dolomite adjacent to the ore bodies, however, suggests that the actual maximum formation temperature of the sphalerite could not have been much in excess of 600° C. Although exsolved chalcopyrite is present in two of these samples, copper analyses indicate that it could not constitute more than two percent of the sphalerites. According to Kullerud (written communication) such small amounts of chalcopyrite do not appreciably effect temperature estimates based on the FeS content of sphalerite, and the author sees no reason to reject them. The temperature estimate of 380° to 405° C based on the iron content of the third sample, which contains no chalcopyrite, is in reasonable accord with the temperature of exsolution of chalcopyrite from sphalerite (350°–400° C) observed by Buerger (3).

Pyrite Geothermometer.—The pyrite geothermometer, developed by F. G. Smith (23), is based on the theory that the thermo-electric potential of pyrite crystals varies inversely with their temperature of formation. Smith points out that clean polished surfaces of single crystals should be used for temperature determination and that care should be taken to make sure that neither of the probes of the apparatus is touching an inclusion of some other mineral in the pyrite.

TABLE 2

ZnS AND FeS CONTENT AND ESTIMATED TEMPERATURES OF FORMATION OF SPHALERITE SAMPLES FROM GILMAN

Field	Wt. % calculated		Wt. % corrected[2]		Minimum formation temp. °C	
No.[1]	FeS	ZnS	FeS	ZnS	1 atm.	1,000 atms.
61 T 40	16.91	79.92	17.46	82.54	565	590
12 T 41	10.11	87.43	10.36	89.64	380	405
41 T 41	14.84	83.28	15.12	84.88	510	535

[1] Laboratory number and location given in appendix.
[2] Recalculated to 100% FeS + ZnS.

Field relations and petrographic studies indicate that three main generations of pyrite were formed in the ore deposits at Gilman. The first two, which account for most of the pyrite in the district, had largely ceased to form before the deposition of the base metal sulfide ores began; the third was later than most, if not all of the primary ore minerals. The two older generations of pyrite are commonly fine-grained and intimately associated with each other; the older of the two is characterized by numerous minute inclusions of quartz and dolomite. The third generation of pyrite forms veinlets, which cut the ore sulfide minerals, and coarse crystals, relatively free of inclusions, lining voids and channelways in the ore bodies. Some pyrite was penecontemporaneous with the major sulfide ores but studies of polished ore specimens indicate that it was relatively rare and quantitatively unimportant.

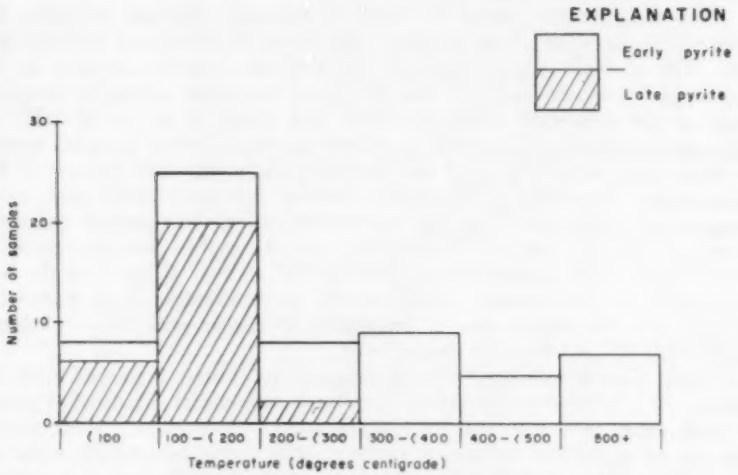

FIG. 3. Temperatures of formation of pyrite samples from Gilman as determined with the pyrite geothermometer.

Sixty-two polished specimens of pyrite from Gilman were tested with the pyrite geothermometer and four or five readings were made on each specimen; most of these replicate readings showed a range of less than 100° C. The mean value of the readings for each specimen was then recorded. The results are summarized in Figure 3. Nearly all of the samples tested were too fine-grained to permit the placing of both probes on a single crystal, and many of them showed evidence of more than one generation of pyrite on a single polished surface. A further source of error is the presence of numerous minute inclusions of dolomite, quartz and sulfide minerals, invisible to the naked eye, in much of the pyrite. The bars on the histogram (Fig. 3) may therefore represent many readings that were invalid.

The restricted range of the late pyrite with a modal concentration in the

100°–200° C class, contrasts sharply with the extended range of the early pyrite with a modal concentration in the 300°–400° C class. This suggests that most of the late pyrite was deposited at a lower temperature than most of the early pyrite. According to Thode, MacNamara, and Collins (26, p. 369), H. S. Scott reported formation temperatures of 235° C and 185° C respectively on coarse and fine-grained pyrite associated with chalcopyrite in a specimen from the Gilman district. Scott's estimates are also based on readings from the pyrite geothermometer. The association of pyrite with chalcopyrite rather than iron-rich sphalerite suggests that both pyrite samples tested by Scott were probably late pyrite. The possibilities for error inherent in the application of the pyrite geothermometer to pyrite samples from the Gilman district make it difficult to draw any firm conclusions from the resulting data. The high proportion of early pyrite samples that yielded temperature estimates below 500 degrees is difficult to reconcile with the evidence that the high iron sphalerite, which formed later, probably crystallized at temperatures close to or above 500°. This can be interpreted either as evidence that the temperature readings on the early pyrite are invalid, or as evidence that the early pyrite was deposited chiefly during an initial barren "heating up" stage and had largely ceased to form before the advent of the main ore stage when temperatures were dropping.

Fluid Inclusions.—If samples containing two-phase fluid inclusions are slowly heated on a microscope stage until the gas bubbles disappear the temperature at which they disappear, corrected for confining pressure, furnishes an estimate of the temperature at which the mineral containing the inclusions was formed. This is one of the oldest and most widely used methods of geothermometry. It was first proposed by H. C. Sorby (24) who stated the underlying assumptions and applied the method to inclusions in quartz crystals heated in a paraffin bath. More recently Ingerson (9) and Kennedy (10) have discussed the technique and given pressure-temperature correction curves.

Hydrothermally altered rocks in the vicinity of the ore deposits were selected for study. Chips containing fluid inclusions were centered on a microscope heating stage, and the mean temperature at which bubbles in the inclusions disappeared was recorded when possible. Unfortunately the equipment used for this study was limited to temperatures below 300° C and many of the bubbles failed to disappear at this temperature. The samples studied and the observed filling temperatures of their fluid inclusions are listed in Table 3.

The assumption was made that these were primary, not secondary, inclusions because they were randomly distributed through the crystals rather than concentrated along fractures and grain boundaries. The possibility of leakage into or out of the crystals since their formation was discounted since the bubbles in these inclusions returned to their original size and shape when the samples were cooled. Unfortunately the inclusions were too small and sparse to permit analysis of the fluids contained, so that the assumption of H_2O content rather than CO_2 rests solely on the fact that the bubbles showed no tendency to expand during heating.

These temperatures uncorrected for pressure represent minimum estimates. The appropriate pressure correction would raise these temperatures by 20° to 40° C, and will be considered later.

The hydrothermal dolomitization of carbonate rocks in the Gilman area may be subdivided into three stages. The earliest stage converted the Leadville limestone into a dark fine-grained dolomite over an area of many square miles. The second stage was more limited in extent; it produced the zebra banded dolomite (nos. 3, 4 and 5, Table 3), which is largely confined to the upper part of the Leadville limestone in the vicinity of the ore deposits, although some is found in the lower part of the Leadville and the Dyer dolomite member of the Chaffee formation. The last stage is confined to a zone one to four feet thick about 95 feet above the base of the Leadville within the

TABLE 3

FILLING TEMPERATURES OF FLUID INCLUSIONS IN SAMPLES FROM THE GILMAN DISTRICT

No.	Description	Mean filling temp. degrees C.
1.	Undolomitized recrystallized Leadville limestone, 5 miles from ore deposits.	180°
2.	Pink hydrothermal dolomite, midway between 16–17 manto orebody and 35 chimney orebody, about 400 ft. from ore, Eagle mine.	220°
3.	Altered Dyer dolomite member of the Chaffee formation, about 40 ft. under upper end No. 1 manto orebody, 15 level Chief mine.	More than 300°
4.	So-called Zebra dolomite N. of tailings pond, 2 miles from ore deposit.	Do.
5.	So-called Zebra dolomite replaced by coarse dolomite, about 1,000 ft. N. of No. 1 orebody, Eagle mine.	do.
6.	Quartzite lens filling channel in altered Leadville limestone about 800 ft. from ore, Eagle mine.	275°

mineralized area. This zone has been brecciated, and consists of fragments and larger blocks of a light pink to buff dolomite in a matrix of dark hydrothermal clay (no. 2, Table 3).

The sample data in Table 3 suggest that the maximum rock temperatures during dolomitization were above 300° C, but that temperatures had declined to slightly above 200 degrees before dolomitization ended. If these samples are reasonably representative of the rock types from which they are taken, it appears that the temperature in the undolomitized limestone (sample 1) reached a maximum of about 200° C. Unfortunately the early dark dolomite samples are uniformly fine-grained and no fluid inclusions with bubbles could be observed in them; hence it was impossible to obtain temperature estimates on them by this method. Samples representing the ensuing zebra dolomite alteration contain inclusions that fill above 300° C, but one of the specimens is near a hydrothermal conduit in which an ore body formed later. Inclusions

in a sample of the youngest pink dolomite (sample 2) show a decline in filling temperature to about 200° C.

Thermoluminescence.—Many minerals, including the carbonates, release light energy when they are heated at temperatures well below that of incandescence. This property of thermoluminescence has been studied by a number of workers; among the most thorough and comprehensive of recent thermoluminescence investigations are those conducted by Farrington Daniels, D. F. Saunders and their associates at the University of Wisconsin. Thermoluminescence apparatus consists essentially of a heating stage, photomultiplier tube, and recorders for intensity of heat and light. If the light emitted is plotted as a function of temperature of the sample, the resulting "glow curve" commonly shows several "glow peaks" of maximum intensity, at temperatures that are fairly consistent for a particular mineral. If a sample is heated, cooled, and then reheated, the glow peaks, at temperatures below that to which it was originally heated are largely destroyed. Daniels and Saunders (written communication) conducted experiments on limestone samples taken close to

TABLE 4

THERMOLUMINESCENCE AND FLUID INCLUSION DATA ON LIMESTONE AND HYDROTHERMAL DOLOMITE FROM THE GILMAN DISTRICT

No.	Thermoluminescence glow peaks[1]		Fluid inclusion filling temperature
	235° peak	330° peak	
1	moderate	strong	180°
2	weak	strong	220°
3	absent	very weak	more than 300°
4	absent	strong	Do.
5	absent	strong	do.

[1] Glow curves run by A. P. Pierce.

contacts with igneous intrusions, and concluded that the contact temperatures could be estimated, within limits, by noting the points on the glow curves below which the characteristic peaks have been suppressed.

Thermoluminescence glow curves were obtained for the same five dolomite and limestone samples used in the investigation of fluid inclusions, to test the assumption that their formation temperature represented the maximum temperature to which they had been subjected, and that glow peaks below that temperature would have been suppressed. The thermoluminescence apparatus available was designed for use only in the 200–400 degree range. Undolomitized limestone (sample 1) shows well defined peaks, near 235° and near 330° C (Table 4). In the other specimens, however, the 234° C peak was suppressed, yet in all but one of these specimens the 330° C peak is well preserved. The sample in which the 330 degree peak is also suppressed was taken within 40 feet of a major ore body and may well have been reheated by the passage of ore solutions to a temperature considerably above that at which it formed.

There is some range in the temperature, and even more in the amplitude, of glow peaks from different samples of calcite, and the range between calcite and dolomite is still greater. The presence of a weak 235 degree peak in a single hydrothermal dolomite sample, (no. 2) from the Gilman district, suggests that, in this area, there is some correspondence in glow peak temperatures between the two minerals, but this may not be true of other areas of dolomitized limestone. Even at Gilman, the correlation between suppression of glow peaks and fluid inclusion temperatures, shown in Table 4, is based on a small number of samples and may be fortuitous. A considerably larger representative suite of hydrothermal dolomite samples must be checked by this method before its worth as a geothermometer can be properly evaluated.

The temperature estimates given by the two methods are reasonably consistent, thus adding support to the theory that the hydrothermal dolomitization was largely accomplished in the temperature range between 220° C and 330° C.

DEPTH AND PRESSURE ESTIMATES

Four lines of evidence have been used to estimate the depth at which the Gilman ore deposits formed: (1) reconstruction of the stratigraphic column at the time of ore deposition; (2) extrapolation of the geothermal gradient; (3) the pressure correction necessary to reconcile thermoluminescence data with fluid inclusion data on the dolomite samples; and (4) the probable crushing strength of the dolomite surrounding the open pre-ore channels.

Reconstruction of the stratigraphic column that existed at the time of mineralization, probably during the early Tertiary period, furnishes the most direct approach to the depth problem. In areas where the paleogeography is well understood, and the time of mineralization can be closely estimated, the limits established by this method are close enough for all practical purposes. Unfortunately, in the Gilman area the Mesozoic sedimentary rocks have been completely eroded and the extent to which this erosion had progressed at the time of ore deposition cannot be determined. Furthermore, there is no direct evidence as to the time of mineralization and the indirect evidence places rather broad limits. A minimum depth can be estimated from the thickness of the rocks of Late Paleozoic age now remaining in the area. These rocks above the Leadville limestone have a maximum thickness of approximately 8,000 feet in the Pando area a few miles southeast of Gilman (27). The minimum cover, from stratigraphic evidence, is therefore approximately 1½ miles. This thickness corresponds to a lithostatic pressure of about 650 kilograms per square centimeter. The maximum cover that could have existed when the ore deposits were formed is more difficult to estimate. Nearly 5,000 feet of Mesozoic sedimentary rocks are exposed in the valley of the Blue River about 25 miles northeast of Gilman (14), and a somewhat thicker section occurs a similar distance to the northwest between Wolcott and State Bridge (Lovering, T. S., oral communication). An isopach map of the Upper Cretaceous rocks of this region compiled by Reeside (20) indicates a total thickness of these rocks in the Minturn area of between 5,000 and 6,000 feet. Approximately 500 feet of Triassic and Jurassic sedimentary rocks are

present on Red and White Mountain, fifteen miles north-northwest of Gilman, between the Maroon formation of Pennsylvanian and Permian age and the Dakota sandstone of Cretaceous age (16, p. 49–51). This evidence suggests that the Leadville limestone at Gilman may have been buried beneath as much as 15,000 feet of rock when the area became emergent near the end of the Cretaceous period. If erosion was negligible prior to the emplacement of the ore deposits, we can then estimate a maximum depth of about three miles, corresponding to a lithostatic pressure of about 1,200 kg/cm^2 at that time.

Another method for calculating depth of formation of ore deposits consists of extrapolating an assumed geothermal gradient to a depth that would produce a temperature equal to the lowest temperature determined for minerals formed during mineralization. This method involves two large factors of uncertainty, the geothermal gradient assumed and the estimation of minimum temperature. In most hydrothermal ore deposits—and Gilman is no exception—the minimum temperature during mineralization cannot be precisely determined. More than 200 temperature gradients from deep borings in various parts of the United States have been compiled by Spicer (25). A frequency distribution of Spicer's gradients gives a modal value close to 35 degrees C/km or about 1 degree C/100 ft; it also shows a range from less than one fourth to more than three times that figure. If we accept 1 degree/100 ft as a conservative figure, assume a soil temperature of 20° C, and extrapolate to the 150°C estimate of formation temperature for hessite bearing galena, the geothermal gradient method gives a miximum depth estimate of 13,000 feet or about two and a half miles.

The effect of confining pressure on the filling temperature of two-phase fluid inclusions has been calculated by Ingerson (9), and refined by Kennedy (10). The discrepancy between observed filling temperature, at atmospheric pressure, and the actual temperature of formation increases systematically with increasing pressure. If it is possible to obtain an independent estimate of the temperature of formation unaffected by pressure, on a sample with fluid inclusions that fill at a known temperature, the confining pressure at the time of formation can be calculated. Sample no. 2, Table 3, furnishes the data for such a calculation, if certain basic assumptions are granted. These assumptions are: first, that the destruction of thermoluminescence glow peaks is a function of temperature only, and is unaffected by pressure; second, that the temperature of formation of the sample was not exceeded in its subsequent geologic history; third, that the partial destruction of a glow peak indicates that the temperature of this peak was almost reached but not exceeded; fourth, that consistent filling temperatures of the fluid inclusions are reliable estimates of formation temperature uncorrected for pressure; and finally, that calcite and hydrothermal dolomite from this area give glow peaks at similar temperatures. The first assumption is based on the studies of Zeller (29), who found that the amplitude of artificially induced glow peaks in carbonate rocks was decreased by subjecting them to high pressures but that the temperature of these peaks remained essentially unchanged. The second assumption is based on the premise that, had the temperature of formation been subsequently ex-

ceeded, the fluid inclusions in the rock would have leaked out along cleavage planes and grain boundaries. The third assumption is probably the least reliable, because the amplitude of glow peaks is known to be a function of a number of complex interdependent variables, and also because glow peaks, once destroyed, may tend to slowly re-establish themselves with the passage of time. The reasoning back of the third assumption is this—temperatures attending the dolomitization of these rocks at depth might be expected to change very slowly, so that if a glow peak temperature was substantially exceeded, the new temperature would probably be maintained for a sufficient length of time for the complete destruction of that peak; the fact that the 235° C peak was completely destroyed in some of the samples justifies the assumption that the restoration of these peaks with the passage of time, was negligible. The fourth assumption, concerning the validity of fluid inclusion temperatures as geothermometers, seems justified by the results of fluid inclusion studies on synthetic crystals, in which the filling temperatures were within three or four degrees of the true formation temperature (21). The last assumption concerning similarity of glow peaks from calcite with those from dolomite has been previously discussed (p. 702). In dolomite sample no. 2 fluid inclusions filled at 220° C and the 235° C glow peak had been partly destroyed. A pressure correction corresponding to a temperature discrepancy of 15° to 20° C therefore appears reasonable. According to the curves published by Kennedy (10), this pressure correction would correspond to a depth of approximately ¾ of a mile, assuming that the fluid in the inclusions is nearly pure water. This figure is about half of the lower limit inferred from stratigraphic evidence, but is in reasonable accord with the paleo-geomorphology of the region to the east studied by T. S. Lovering (oral communication).

The pressure correction corresponding to the minimum depth of 1½ miles, inferred from stratigraphic evidence, would amount to about 40° C. This would suggest a temperature in the neighborhood of 260° C, for the formation of this dolomite (sample no. 2). This discrepancy does not necessarily invalidate the fluid inclusion-glow curve depth estimate, because the minimum stratigraphic depth was based on the assumption that erosion prior to the emplacement of the ore deposits at Gilman had not cut deep enough to remove any of the Paleozoic rocks. This is an assumption that cannot be proved.

The presence of a number of large, open, pre-ore cavities in the dolomitized limestone at Gilman suggests that the crushing strength of the dolomite might be used to establish a maximum value for the depth at which the ore bodies could have formed, since these cavities were apparently leached out after the dolomitization of the limestone but before the introduction of the ore depositing solutions. This approach to the problem, like the others, is fraught with complications. In the first place there is no direct evidence as to whether the cavities were open, or filled with fluid prior to the introduction of ore solutions, but presumably they were filled; in the second place, the presence of numerous fractures and small faults in the rocks around these cavities suggests that some of the pressure caused by the weight of overlying rocks might have been relieved by minor readjustments along these planes of weakness. Griggs (6, p. 116) gives the mean unconfined crushing strength of limestone as

960 Kg/cm² but recent experiments by Robertson (22) indicate that under hydrostatic confining pressure the crushing strength of limestone from Solenhofen is close to 3,000 Kg/cm². On the other hand, high temperature and mineralizing solutions drastically reduce the crushing strength of carbonate rocks. Griggs (6, p. 114) found that in the presence of carbonated water at a temperature of 150° C the crushing strength of Yule marble was reduced by 80 percent. It thus appears that if the cavities contained liquids—and of this there is little doubt—the rocks would have been weakened due to the temperature and chemical nature of these solutions. A maximum depth estimate based on the unconfined crushing strength should therefore be conservative—if we can assume that the walls of the cavities supported the full weight of the superincumbent rock. If we use the average crushing strength figure given by Griggs for limestone, this would correspond to the lithostatic pressure of approximately 2½ miles of overlying rock. This method of approach to the depth problem would be considerably improved if it could be based on crushing strengths determined on samples taken from the sides of open pre-ore channels in the Leadville limestone at Gilman. Unfortunately the necessary equipment for determining crushing strength of rock samples was not available to the author.

The evidence now available on the pressure and depth at which sulfide ores formed in the Leadville limestone at Gilman is somewhat contradictory. The most we can say is that most of it suggests a depth of not less than one mile nor more than three at the time of mineralization. This points up the need for research directed towards extending and improving the methods of "geobarometry" of ore deposits.

SUMMARY AND CONCLUSIONS

Mineralization of the Leadville limestone in the Gilman district took place in two or more distinct major stages. The first stage, hydrothermal dolomitization, reached a maximum temperature slightly above 300° C, and there is some evidence suggesting that temperatures had declined to about 240° C before it ended. Dolomitization was followed by a period of solution channeling and leaching at indeterminate temperatures. This in turn was followed by primary sulfide mineralization, beginning with deposition of pyrite and manganosiderite, presumably at temperatures of as much as 600° C. The pyrite was followed by sphalerite, which began to form at about 590° C and had largely ceased by the time temperatures had declined below 350° C. Chalcopyrite succeeded the sphalerite at a somewhat lower temperature. If there was only one period of mineralization, the main period of copper mineral deposition ended with the nearly simultaneous formation, locally, of bornite, tetrahedrite, and chalcopyrite below 250° C. After the copper, lead and silver, represented chiefly by galena and hessite, were deposited near or below 150° C. Primary hydrothermal metallization appears to have closed with late pyrite and native gold, some of which may have formed at or below 100° C.

There is a distinct possibility that ore sulfide deposition at Gilman took place in two separate phases. If so, the first phase resulted in the emplace-

ment of the large zinc manto ore bodies, the second was responsible for emplacement of copper-lead-silver ore in the pyritic chimneys. If this were the case the foregoing picture would have to be modified somewhat. The early high-temperature phase, during which porous pyrite formed in the chimneys, and sphalerite was deposited in the mantos, probably closed with the local deposition of non-argentiferous galena in the mantos at a temperature of about 300° C. The late phase of chimney ore deposition opened with sparse pyrrhotite and abundant chalcopyrite forming below 250° C, followed by tetrahedrite, bornite, galena and hessite ending below 150° C, and closed with the deposition of late pyrite and native gold.

The depth at which these events took place is difficult to estimate. Available evidence suggests that it was probably more than 1 mile and less than 3 miles. This in turn would indicate a lithostatic confining pressure between 400 and 1,200 atmospheres.

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APPENDIX

Numbers and locations of samples for which analytical data are reported in Tables 1–4.

Tables 1 and 2

Field No.	Lab. No.	Location
61TL40	243809	Chimney ore, a proturberance of 4–5 orebody, 17–17 level, 100 ft from station on 16–16 incline.
12T41	243806	Manto ore No. 1 orebody, 15 level, 15–13 just above 16–14 ore.
41T41	243808	In channel, 500 ft N. of lower end, No. 1 manto orebody, 18–10, near 18–8.

Tables 3 and 4

No.	Field No.	Location
1	51TL40	Leadville limestone, near fault, N. edge of bench S. of lower Grouse Creek.
2	27T41	Pink breccia, side drift in position of 16–32, Eagle mine.
3	63T41	Altered Dyer dolomite member of Chaffee formation, 15 level, Chief mine.
4	30T46	So-called Zebra dolomite ending at joint, 8,580 ft knob N. of tailings pond.
5	9TL40E	So-called Zebra dolomite, no. 2 crosscut, 16th level, Eagle mine.
6	110T41	Quartzite lense in Leadville, 19–10 level, Eagle mine.

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TITANIFEROUS SEDIMENTARY ROCKS IN THE CUYUNA DISTRICT, CENTRAL MINNESOTA[1]

ROBERT G. SCHMIDT

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ABSTRACT

The ore-yielding main iron-formation of the Cuyuna district is stratigraphically overlain by approximately 300 feet of titaniferous sedimentary rocks, and a field test for titania has been an effective supplement to customary stratigraphic studies. The titanium is in leucoxene.

The sedimentary rocks of the district are highly folded, and the main iron-formation serves as the reference for relative stratigraphic position. The strata older than the main iron-formation are light-gray argillite and siltstone with local lenses of quartzite near the upper contact; the older strata contain about 0.5 percent titania. The main iron-formation is a persistent layer of chemically precipitated ferruginous sediments that contain about 0.2 percent titania. The strata younger than the main iron-formation are gray or black argillite, partly carbonaceous, partly ferruginous, and are transitional into lenses of lean upper iron-formation. At the base of the younger strata, 250–350 feet of argillite and slate contain 1–3.7 percent titania and average 2.0 percent. This titaniferous argillite is similar to argillite higher in the younger sequence which contains less than 1.5 percent titania and averages about 1 percent. There are weathered rocks interpreted as tuffs and basalt flows locally interlayered near the bottom of the titaniferous zone. The basalt contains 1–2 percent titania. The high titania content of the 250–350-foot zone suggests a volcanic origin; the argillite probably is reworked basaltic pyroclastics. A few titaniferous bodies have been found in rocks below, within, and above the main iron-formation and thin titaniferous argillaceous layers occur within the main iron-formation.

INTRODUCTION

THE geology of the North range, Cuyuna iron district, Minnesota (Fig. 1), has been studied by the U. S. Geological Survey from 1951 to 1957. In the

[1] Publication authorized by the Director, U. S. Geological Survey.

course of the work the distribution of titanium in the Precambrian metamorphosed sedimentary and igneous rocks was investigated and this paper is a discussion of the results.

The Cuyuna iron-formations are included in a metamorphosed sequence of dominantly argillaceous sedimentary rocks. Most of the sequence contains

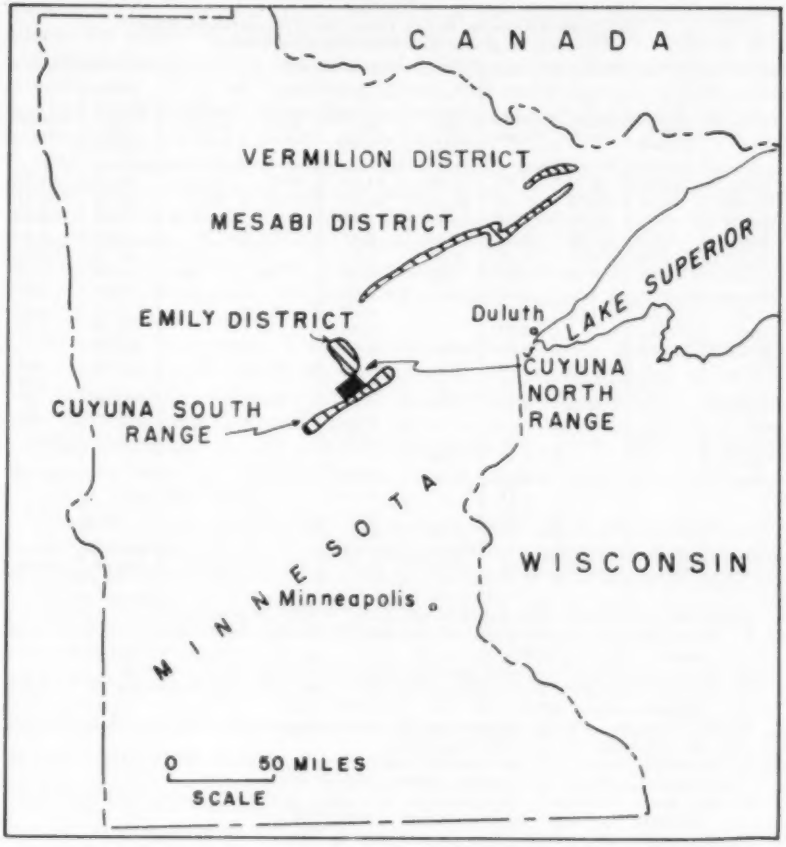

FIG. 1. Index map of Minnesota showing location of the North range, Cuyuna district.

titania in the amounts normal for such rocks: about 0.2 percent titania in the iron-formation and about 0.5 percent titania in the argillite. Argillaceous metasedimentary rocks having abnormally high titania content (1.0–3.7 percent) occur consistently throughout a 250–350-foot stratigraphic zone immediately above the main iron-formation and locally elsewhere in the stratigraphic

section. Other common rocks that contain TiO_2 in excess of 1.0 percent are: (1) layers thought to be basalt flows and water-laid basaltic tuff that in part of the district are intercalated as components of the 250–350-foot titaniferous zone; and (2) completely chloritized intrusives.

TABLE 1

THE COMPOSITION OF ROCKS FROM THE TITANIFEROUS ZONE AND SELECTED ANALYSES FOR COMPARISON

	A Meta-basalt, Cuyuna	B Basalt, Hawaii	C Tuff, Cuyuna	D Black slate, Cuyuna	E Gray argillite, Cuyuna	F Gray slate, Cuyuna	G Black argillite, Cuyuna	H Altered basaltic tuff, Hawaii	I Average shale
SiO_2	48.92	49.73	57.27	37.20	64.06	67.2	43.5	55.58	58.38
Al_2O_3	14.76	13.71	23.76	27.39	20.95	13.1	22.3	19.13	15.47
Fe_2O_3	2.54	2.92	1.63	12.24	1.49	9.1	16.2	6.40	4.03
FeO	8.96	8.64	0.62	2.41	0.42	.45	*	0.50	2.46
MgO	6.78	8.27	1.38	1.00	1.11	.78	.74	0.52	2.45
CaO	11.50	9.10	0.05	.45	0.01	.08	.23	0.44	3.12
Na_2O	1.52	3.16	0.07		0.06	.10	.11	4.59	1.31
K_2O	0.53	1.02	6.43		4.52	4.5	3.7	2.84	3.25
H_2O^-	0.11		0.05		0.03			2.74	1.34
H_2O^+	2.97		4.63	9.14	4.84			5.58	3.68
TiO_2	1.27	2.84	3.38	3.68	1.04	1.8	2.3	1.46	0.65
CO_2	0.02		0.02	.51	0.01				2.64
P_2O_5	0.16	0.48	0.23	.316	0.14	.08	.23	0.10	.17
SO_3					0.05				.65
MnO	0.18	0.13	0.00		0.00	.04	.02	0.31	trace
Mn_2O_3				.23					
BaO			0.16		0.06				0.05
F			0.11		0.12				
S			0.05		0.33				0.81
C				3.75	0.53				
Total	100.22	100.00	99.84	98.32	99.77	97.23	89.33	100.19	100.46

Blank space indicates no determination was made. * FeO not determined.

A. Metabasalt from Maroco mine, Cuyuna district, Minn.; Faye H. Neuerburg, U. S. Geol. Survey, analyst; (C589).

B. Average of 56 basalt analyses, Island of Hawaii (Washington, 1923, p. 361).

C. Tuff bed from Joan mine, Cuyuna district, Minn.; Faye H. Neuerburg, U. S. Geol. Survey, analyst; (C591).

D. Black slate from titaniferous zone near Huntington mine north of Blackhoof Lake (Harder and Johnston, 1918, p. 120, column 14).

E. Gray and black titaniferous argillite from Joan mine, Cuyuna district, Minn.; Faye H. Neuerburg, U. S. Geol. Survey, analyst; (C590).

F. Gray laminated titaniferous slate from Huntington mine, Cuyuna district, Minn.; S. M. Berthold, U. S. Geol. Survey, analyst ;(52-1419CW).

G. Black argillite from Feigh mine, Cuyuna district, Minn., S. M. Berthold, U. S. Geol. Survey, analyst; (52-1423CW).

H. Yellow tuff no. 323, Island of Hawaii (Wentworth, 1938, p. 128).

I. Composite analysis of 78 shales (Clarke, 1924, p. 30).

Titania analyses of Cuyuna rocks were first studied by the Geological Survey to determine if certain oxidized and chloritized ferruginous rocks within the oxidized iron-formation were intrusives. As described by Tyler (6, p. 1108), the titania content is a relatively easy way to distinguish altered intrusives containing 20–30 percent iron from enclosing iron-formation. To

determine the possible application of this method in the Cuyuna district, C. E. Dutton examined a group of analyses of Cuyuna rocks (3, p. 120). He obtained no data pertinent to the recognition of intrusives, but he noted the conspicuously high titania content (3.68 percent) in one of the sedimentary rocks (Table 1, column D). At a field conference in the Cuyuna district in 1952, the above-mentioned titaniferous sedimentary rock was discussed and Messrs. R. A. Ekar and A. C. Borgeson of the Snyder Mining Company offered the results of some of their studies of Cuyuna district rocks to the Geological Survey. They had prepared a large number of complete analyses of drill samples from the iron-formation and wall rocks near the Martin mine and had concluded that certain titanium-rich layers could be correlated from hole to hole.

The Geological Survey made about 700 field analyses to determine the distribution of titanium in Cuyuna rocks, using the method described by Shapiro and Brannock (5). Much helpful assistance was given by many mining companies in the district, particularly the M. A. Hanna Company, the Snyder Mining Company, and the Oliver Iron Mining Division, United States Steel Corporation, who supplied complete analyses and hundreds of selected samples for testing.

As testing progressed, it was evident that most older metasedimentary rocks and the main iron-formation contained the small amounts of titania normal to these types of rocks but that the lowermost 250–350 feet of younger beds near the main iron-formation contained from three to six times the normal amount and formed a distinct titaniferous zone. It was found that the igneous rocks and certain thin sedimentary layers within the main iron-formation were titaniferous.

Metasedimentary rocks older and younger than the main iron-formation in the Cuyuna district can be readily distinguished by inspection of exploration drill samples if the samples are cores or coarse chips of relatively fresh argillite, quartzite, or tuff. Unfortunately, many samples are bleached argillite and provide little information for determination of the stratigraphic position. For such samples the titanium content may be the only information by which the zone immediately above the iron-formation can be distinguished from rocks both younger and older. The titanium test has been successfully used to supplement conventional methods of study in planning a small amount of exploration drilling in parts of the district known as the North and South ranges. Study of the titania analyses of many drill samples was used in preparing maps of the North range.

STRATIGRAPHY

The folded Precambrian metasedimentary rocks of the Cuyuna district are covered with Pleistocene glacial drift to depths of 16 to 200 feet throughout the area studied, and the bedrock can be observed only in mine workings and in drill samples. The iron-formations of the district are included in a sequence of argillite and slate and lesser amounts of coarser clastic and volcanic rocks (Fig. 2) of possible Animikie age, following the usage of Grout et al.

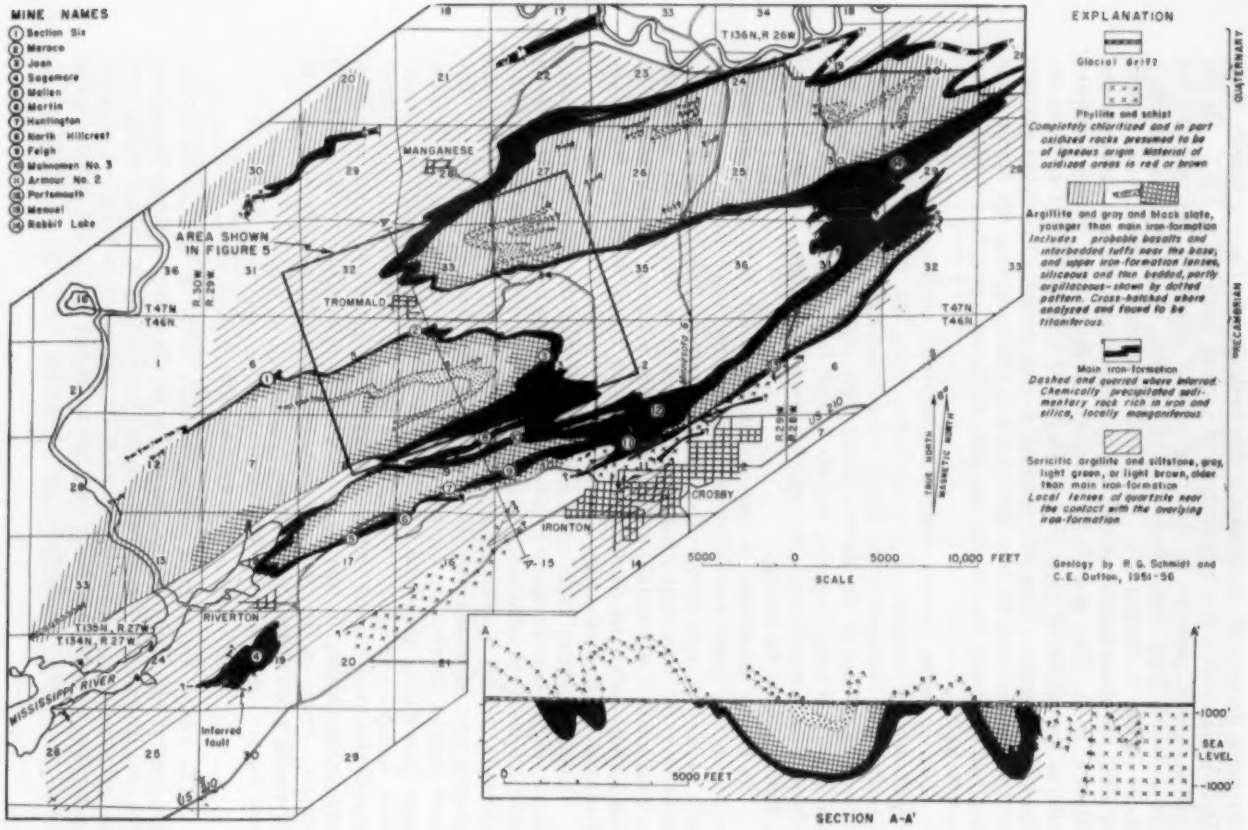

FIG. 2. Bedrock geology of the North range, Cuyuna district, Minnesota.

(2, p. 1021, 1050). The top and bottom of the main iron-formation serve as the reference horizons for stratigraphic measurements (Fig. 2). Below the main iron-formation is an undivided sequence of light- and dark-gray, light-green, and light-brown argillite, slate, and fine siltstone; lenses of quartzite are locally interbedded in the upper part. The titania content of these older metasediments is generally about 0.5 percent, about normal for argillaceous rocks (Table 1, column I). Above the main iron-formation are interbedded argillite, slate, minor siltstone, and one or more siliceous argillaceous iron-formation lenses; interlayered near the base in part of the district are rocks believed to be tuff and basalt. The lowermost 250 to 350 feet of the younger sequence contains 1.0 to 3.7 percent titania, averaging 2.0 percent, whereas the rocks above contain a trace to about 1.5 percent titania, averaging about 1.0 percent. The argillite and slate of the younger sequence are in part distinctly laminated with ten or more individual layers per inch of thickness, in part massive bedded in layers up to several feet thick. Either variety may be gray, black, green, brown, red, or nearly white—the first three are original colors, and the last three products of oxidation and leaching. The nature and the amount of the black carbonaceous matter in the argillite and slate are not known but the metamorphism probably has not been sufficiently intense to produce graphite. Changes to light colors along cracks and bedding planes are common.

Iron-formations

Main Iron-formation.—The main iron-formation is a persistent unit of chemically precipitated sedimentary rock rich in silica and iron, containing variable amounts of manganese and generally sparse amounts of clastic material. Almost all the iron and manganiferous iron ores produced in the district are found in this unit.

The main iron-formation has been divided into two mappable types, the thick-bedded and the thin-bedded facies. In part of the district the entire iron-formation is thick-bedded, in part it is all thin-bedded, and in about one-third of the area the thick-bedded overlaps the thin-bedded facies and grades downward into it. The iron minerals in the unoxidized parts of the thin-bedded facies are dominantly silicates, siderite, and magnetite, and in the fresh thick-bedded facies, hematite, goethite, sparse magnetite, silicates, and siderite; in oxidized and leached parts of both facies little else than ferric-iron oxides and quartz remain. The thickness of the iron-formation changes gradually from less than 45 feet to 500 feet or more. Relatively thin layers of titaniferous rock are interbedded in the iron-formation in several places.

Upper Iron-formation.—Parts of the younger argillite and slate contain an average of as much as 20 percent iron, and they grade into lenses of siliceous argillaceous thin-bedded iron-formation several hundred feet thick, called upper iron-formation. The original iron mineral was probably finely divided pyrite, which remains in part of the rock, and perhaps siderite in some parts that now contain red and brown oxides. Except for some thin lenses, which are 10 feet or less thick, the upper iron-formation occurs in lenses in the younger

argillite from about 500–700 feet to perhaps 2,000 feet above the main iron-formation; the sequence above is not known. The upper iron-formation is relatively unproductive.

The upper iron-formation is similar to the thin-bedded facies of the main iron-formation except that favorable conditions for deposition must have prevailed over smaller areas, the beginnings and ends of the favorable periods came more gradually, and organic remains and argillaceous sediments were deposited more abundantly with the iron-rich chemical sediments.

Titaniferous Argillites Within the Main Iron-formation.—In the Sagamore mine a bed of argillaceous rock ("paint rock"), perhaps 20 feet thick, is intercalated near the center of the main iron-formation. Parts of this intermediate argillite layer contain 1.0 to 2.0 percent titania. Textures of the argillite at two places in the mine resemble those associated with decomposed tuffs above the main iron-formation in other parts of the district. The writer believes that part of the intermediate argillite is reworked volcanic ash and is similar in origin to the rocks in the titaniferous zone above the iron-formation. Perhaps the less titaniferous parts represent gradations to the normal enclosing iron-formation.

In the Feigh, Mahnomen No. 3, Portsmouth, Manuel, and Rabbit Lake mines, dozens of small concordant layers of titaniferous ferruginous rock have been found within the thin-bedded facies of the main iron-formation. Though probably all related in origin, these rocks are variously schistose or slaty, and some contain rounded or angular chert fragments. The layers are exposed in narrow bands about one foot to 20 feet wide and can be traced perhaps 50 to 100 feet along the strike. The small size of exposures limited tracing these layers to their full extent, but some exposures were long enough to prove that the layers pinch and swell and probably disappear along strike, and that some of the lithologic characteristics change along strike. The layers generally conform to the bedding of the iron-formation, but cross-cutting of several inches of beds, probably the result of local channeling and subsequent filling by a water-borne sediment (Fig. 3), could be seen along a very few contacts. The most common type of rock in these intercalated layers is relatively fine-grained chert breccia or conglomerate; most of the chert fragments are less than one inch in greatest dimension. The chert fragments exceed half the volume of very few of the layers and most commonly comprise less than one-tenth of the volume; the breccias grade into dense slaty rocks without chert fragments. Where the layers are enclosed by oxidized iron-formation, the chert fragments are white and softened, the interstitial material is dark red-brown, ferruginous, and probably argillaceous. Where the enclosing iron-formation is unoxidized, the chert is light to dark gray-green and the interstitial material is hard and black. Some of the concordant bodies are gray-green or red-brown slates or phyllites. Several of these in the Portsmouth, Feigh, and Mahnomen No. 3 mines contained abundant tourmaline in disseminated needles or braided patterns of veinlets. The largest mass of titaniferous rock within the main iron-formation is in the Portsmouth mine, in the NE¼NW¼, sec. 11, T. 46 N., R. 29 W., and is more than 500 feet long and

FIG. 3. Titaniferous chert breccia cutting across bedding of unoxidized thin-bedded iron-formation, Portsmouth mine. The breccia is probably a channel filling. Only the matrix of the breccia is titaniferous.

FIG. 4. Tuffaceous layer overlying a light-colored bed of titaniferous argillite, Maroco mine. Note gradation in grain size in upper bed.

40 to 60 feet wide. The rock is gray schist, with slabs of iron-formation as much as a foot long included near the contacts. The general argillaceous appearance, the chert breccias, and the lack of metamorphic effects at the contacts suggest that these layers are contemporary sedimentary rocks rather than fine-grained sills.

The Titaniferous Zone

The layered rocks in the 250–350 feet above the main iron-formation are titaniferous. These metasedimentary rocks are dominantly gray and black argillite, but locally they include some highly decomposed rocks that were probably volcanic in origin. In all characteristics except the titania content and the local interbedded tuffaceous and basaltic layers, the argillite of this zone seems to be similar to the argillite higher in the sequence. The titania content of argillite in the titaniferous zone is strikingly uniform throughout all layers in the zone in most parts of the district. The titaniferous argillite and slate contain 1.0 to 3.7 percent titania (Table 1, columns D, E, F, G), but most analyses are close to 2.0 percent.

The range in measured thickness of the titaniferous zone is probably due in part to differences in stratigraphic thickness and in part to inaccuracy of measurement caused by drag folding and variations in the dip of beds. The top of the zone is transitional, and the titania content decreases upward from about 2.0 percent to less than 1.0 percent through a stratigraphic thickness of about 100 feet. The arbitrary cut-off for the top of the zone was chosen as 1.5 percent.

Gaps in the titaniferous zone on Figure 2 are in places where no samples were available for analysis. Only in secs. 29 and 30, T. 47 N., R. 29 W., was the titaniferous zone sought in drilling samples and not found; however, the stratigraphic sequence is not well understood at this place.

Volcanic Rocks Within the Titaniferous Zone

Several kinds of concordant, generally weathered or otherwise thoroughly decomposed rocks occur locally within the titaniferous zone in the Cuyuna district. These are believed to have been basaltic flows and tuffs, and this interpretation bears directly on the theory of origin of the high titania in the widespread titaniferous zone above the main iron-formation and in the argillites locally interlayered in the main iron-formation.

Concordant basaltic rocks directly overlying or a short distance above the main iron-formation are found over part of the North range. The greatest thicknesses have been measured in the Maroco mine, where the layers are known to be over 250 feet thick and may be much more. The same basaltic layer occurs continuously southwestward to the Section Six mine and southeastward to the Joan mine. In many other places in the district, as in the Mallen, Huntington, Armour No. 2, and Manuel mines, there are sheared and decomposed rocks stratigraphically above the main iron-formation that probably were basalts or massive tuffs but definite identification is now impossible. It is not known if the inclusion of a considerable thickness of basaltic rock in the titaniferous zone increases the total thickness of the zone. The basalts contain 1.0–2.0 percent titania (Table 1, column A).

The basaltic rock exposed in the open pits is weathered to saprolite. Close-spaced joints have aided the decomposition. A small amount of unweathered rock in the centers of jointed blocks is completely chlorotized and consists of a fine matted intergrowth of chlorite-type mineral grains, probably in part

penninite, and lesser amounts of clinozoisite, calcite, and leucoxene. Clinozoisite and leucoxene have roughly preserved the outlines of part of the original mineral grains and indicate that part of the original rock was in general uniformly aphanitic; in a few specimens phenocrysts 0.5–1.0 mm across were developed. Calcite grains are generally porphyroblastic and as much as 3 mm long. The chemical composition of one sample, Table 1, column A, is similar to that of Hawaiian basalts, column B, except for the greater amount of water in the chloritized Cuyuna rock. The difference in titania content is within the range found in basalts from different localities (1, p. 460), and several field analyses of Cuyuna basalts indicated the titania content ranged from 1 to 2 percent. These basaltic rocks are either flows or concordant sills. There is no metamorphism at contacts. Extrusion or intrusion took place in three or more stages as there are thin lenses of sedimentary rock interlayered at two horizons in the Maroco mine, though not exposed elsewhere. The basalts are conformable with the bedding on fold axes, and, if intrusive, were injected before or at an early stage of folding, in contrast to the later dikes and sills in the southeast part of the district that are generally tabular in the plane of the fold axes. After consideration of the concordance of the basalts over an area several miles square, the writer believes that they were flows.

The tuffs are more widely distributed than the basalts, though where most abundant they make up only a small part of the titaniferous zone. The identification of these rocks as tuffs is tentative as no positive criteria have been found so far. The rocks are well-sorted water-deposited sediments. The grain size ranges from coarse, one-half inch fragments down to that of the finely laminated argillites with which they are intercalated. Most of the prominent beds and lenses that have been studied have grains the size of sand or grit, as beds composed of the larger fragments are uncommon. Many layers have well-developed graded bedding (Fig. 4). No cross-bedding has been found.

A significant peculiarity of the tuff layers is the absence of clastic grains of quartz in beds of sand-sized grains. By contrast, the largest grains in "normal" sediments above and below the tuffs are quartz sand. The tuff grains are generally somewhat flattened and elliptical or spindle-shaped in cross-section; and are composed of clay minerals, probably kaolin and muscovite, and minor amounts of orthoclase, quartz, and leucoxene. The grain outlines are most easily seen in hand specimens and only a few can be recognized under the microscope. No grains of the shapes that are considered to be characteristic of tuffaceous materials (4) were seen in thin sections, although it is probable that all glassy fragments would be devitrified in these old rocks. The titania, which ranges from 1 to almost 4 percent of these layers, is present as very fine specks and irregular masses of leucoxene, which, in one sample, was shown to be anatase by studies with an X-ray diffractometer. The specks and masses of leucoxene are more abundant in some grains than in others but are found disseminated through all.

The complete absence of clastic quartz grains in the well-sorted sand-size and coarser layers and the titaniferous clay of which the particles are now

composed make it probable that the grains were derived from a basaltic rock and that deposition took place where the supplies of clastic materials from other sources were relatively small. Perhaps the coarser grained beds give the most information about the source of these rocks. The half-inch fragments may have been pebbles derived from weathering of a terrestrial basalt or pyroclastic fragments derived more or less locally. The preponderance of fine argillites and chemical sediments in the stratigraphic succession suggests low-lying adjacent land masses and sluggish river systems unable to transport pebbles great distances. The writer considers it most probable that basaltic pyroclastic materials were derived from a local source and were sorted and perhaps somewhat rounded by the waters in which they were deposited.

Titaniferous Rocks in Other Parts of the Stratigraphic Succession

Minor amounts of titaniferous rocks were found in all parts of the stratigraphic sequence. Titaniferous argillite was found in very few places in the older metasedimentary rocks: in sec. 6, T. 46 N., R. 29 W., two drill holes penetrated 30 feet of titaniferous argillaceous rock immediately below the iron-formation; in sec. 9, T. 46 N., R. 29 W., a ferruginous sandy titaniferous argillite was exposed at the base of the iron-formation in the N. Hillcrest mine; near the center of sec. 34, T. 47 N., R. 29 W. (see Fig. 5), titaniferous rock, probably slate, was penetrated by one drill hole. Titaniferous argillite above the titaniferous zone in the younger sequence has been found in two places: about 800 feet stratigraphically above the main iron-formation in the SE¼, Sec. 5, T. 46 N., R. 29 W. (see Fig. 5); and 1,500–2,000 feet stratigraphically above the main iron-formation in two drill holes in the NW¼, sec. 25, T. 47 N., R. 29 W. Ferruginous slate, schist, and chert conglomerate or breccia layers, in part titaniferous, are found in the main iron-formation in many places; these were described earlier in this paper.

Chloritized intrusives and phyllites and schists of unknown origin make up the largest number of titaniferous samples obtained by drilling in rocks outside the titaniferous zone. Some of them might be confused with sheared or decomposed basalts above the main iron-formation but they can be readily distinguished from argillites of any age. Chloritized and partly sheared gabbroic intrusive rocks are abundant along the southeast edge of the district in all parts of the stratigraphic sequence; the intrusives contain 1.0–2.0 percent titania. Titaniferous phyllite and schist of unknown origin within the area underlain by older argillite were found in many drill holes near the Rabbit Lake mine, sec. 29, T. 47 N., R. 28 W., and in a few holes in the SE¼SE¼, sec. 31, and SW¼, sec. 32, T. 47 N., R. 28 W.

ORIGIN OF THE TITANIFEROUS METASEDIMENTS

No direct evidence has been found for the origin of the titania in the titaniferous zone and in the argillaceous layers within the main iron-formation. No evidence suggests the titania was other than a constituent of the original sediment. The effects of the weak regional metamorphism are found equally in all parts of the stratigraphic section, and where titaniferous argillite has

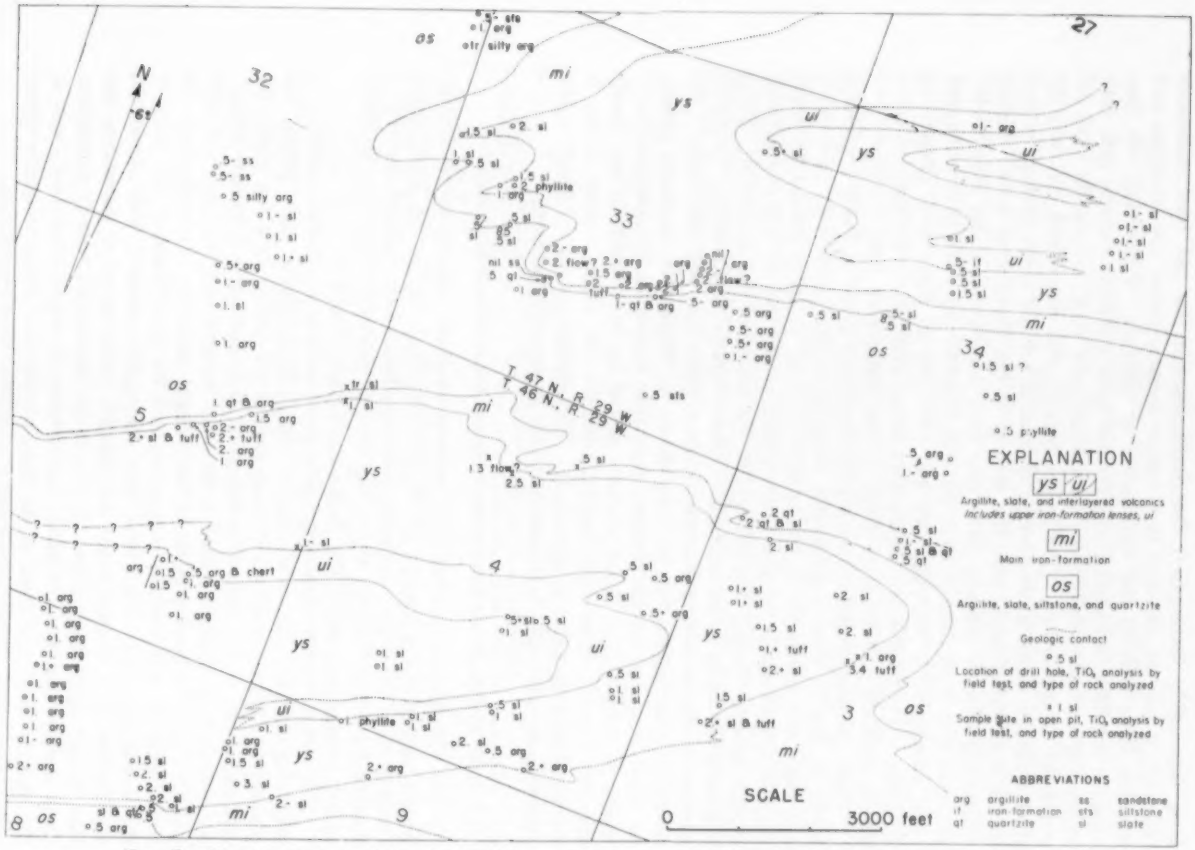

FIG. 5. Titania content of the Precambrian rocks in a part of the North range, Cuyuna district.

been thoroughly decomposed the titania content has not been noticeably changed; it may be assumed that neither metamorphism nor oxidation and leaching have affected the distribution of titanium. Presumably the titaniferous argillaceous material was derived from a different source area than succeeding normally titaniferous metasedimentary rock, despite the close resemblance in appearance. The writer believes that titania-rich clay was derived from the subaerial weathering of basaltic volcanic rocks perhaps related to the tuffs intercalated near the bottom of the titaniferous zone. Slowly weathering basaltic volcanic rocks may have contributed a clay even higher in titania than fresh volcanic rock, and this enriched clay could have undergone some dilution with "normal" clay and yet yield an abnormally titaniferous sediment. The titaniferous argillaceous layers within the main iron-formation almost certainly had the same origin as the titaniferous zone.

Four analyses of titaniferous argillite and slate given in Table 1, columns D, E, F, and G, show considerable variation in the proportions of alumina and silica and variations in the amount of iron, probably caused by slight changes in the relative rates of sedimentation of clay and silt and precipitation of iron minerals rather than changes in the source of the sediments. There is general similarity in the amounts of other chemical constituents. Complete analyses of the succeeding less titaniferous argillite have not been made. The composition of basaltic tuff has been shown by Wentworth to change greatly on weathering, the specific changes depending in part on the seasonal distribution of rainfall (8). Most analyses quoted by Wentworth do not closely resemble Cuyuna titaniferous argillites, but the one which is most similar is reprinted here in Table 1, column H. The greater amount of Na_2O in the Hawaiian sample might be explained by the degree of weathering.

THE USE OF TITANIA ANALYSES IN STRATIGRAPHIC STUDIES

The titaniferous zone has been mapped wherever possible in the Cuyuna North range and the position of the zone has been used to aid stratigraphic correlation, to determine the approximate position of the bedrock outcrop of the iron-formation beneath the glacial drift, to determine order of succession of beds, and to help define the shapes of geologic structures.

The determination of stratigraphic sequence was a common problem in preparing detailed structural maps of the district because tight folding has resulted in high dip angles and many overturned beds. In a few places the rocks adjacent to the main iron-formation lack distinctive lithologic characteristics such as quartzite lenses in the older metasedimentary rock, and tuffs and dark-gray or black argillite in the younger sequence. In such places the titania analysis was the only means available for identification. All parts of the stratigraphic succession have been thoroughly sampled, and titaniferous argillite and slate in exceptional stratigraphic positions are so uncommon that they do not significantly affect the use of the titaniferous zone in correlation. Only samples that were definitely sedimentary in origin were used; most of the igneous rocks and some of the local phyllite and schist of undetermined origin, regardless of their stratigraphic positions, were titaniferous.

CONCLUSIONS

A 300-foot sequence of argillite, slate, very minor siltstone, and, locally, "tuff" and "basalt," immediately above the main iron-formation and certain thin argillaceous layers within the main iron-formation—unlike most argillaceous rocks in the Cuyuna district—contain 2 to 6 times as much TiO_2 as is normal for argillaceous rocks in general. The titania is thought to be derived from weathered basaltic rock and from basaltic pyroclastic material that was incorporated directly into the sediment. Irrespective of origin, the 300-foot thick titaniferous zone has been successfully used as an aid in stratigraphic correlation and in planning the location of exploratory drill holes. A rapid, inexpensive field test was used for TiO_2 analyses (5). These analyses can be used in the proper circumstances to supplement other stratigraphic studies in determining the approximate bedrock position of the iron-formation, the order of succession of beds, and the shapes of geologic structures. Presence of other titaniferous rocks in the sequence, nearly all recognizable by their lithology, makes it absolutely necessary to examine all samples carefully and use only argillaceous metasedimentary rocks for analysis. The titanium test has been found a valuable aid to other methods of study but not infallible when used alone.

U. S. GEOLOGICAL SURVEY,
BELTSVILLE, MARYLAND,
Dec. 16, 1957

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LIMESTONE AQUIFERS OF MARYLAND[1]

E. G. OTTON AND C. A. RICHARDSON

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ABSTRACT

Limestone rocks are an important source of ground water in the Piedmont and Appalachian areas of Maryland. The major limestone aquifers are the Cockeysville and Wakefield marbles and the Silver Run, Tomstown, Frederick, Grove, Waynesboro, Elbrook, Conococheague, Beekmantown, Stones River, Tonoloway, Helderberg and Greenbrier formations. Drilled and dug wells and springs are common sources of water supply in the limestone areas. The depths of 298 limestone wells range from 8.7 to 1,800 feet and average about 150 feet. The yields of these wells range from 0.1 to 575 gpm. Specific capacities, or yield per foot of drawdown, range in 137 wells from < 0.1 to 43 gpm per foot. Springs are common in the limestone areas and their flow varies widely at different times. Most of the known springs are of the 5th magnitude or lower, and therefore have a mean flow of less than 100 gpm. The quality of water from the limestone aquifers is satisfactory for most purposes, but is commonly hard and slightly alkaline. The limestone aquifers are large essentially untapped ground-water reservoirs whose hydrologic character is only partially understood.

GENERAL STATEMENT

LIMESTONE rocks of Precambrian and early Paleozoic age constitute important sources of ground water in the Piedmont and Appalachian areas of Maryland (Fig. 1). In the areas underlain by these rocks, nearly all the residents located beyond the limits of city water mains depend on wells or springs as a source of water supply.

Limestone terranes are those natural regions or areas that are underlain by dominantly calcareous rocks and whose surface and subsurface features are determined by the characteristic response of these rocks to erosion by solution as well as to chemical and mechanical weathering. It is interesting that some areas in the Piedmont of Maryland are underlain by marbles or highly metamorphosed limestones but fail to show many of the typical solutional features

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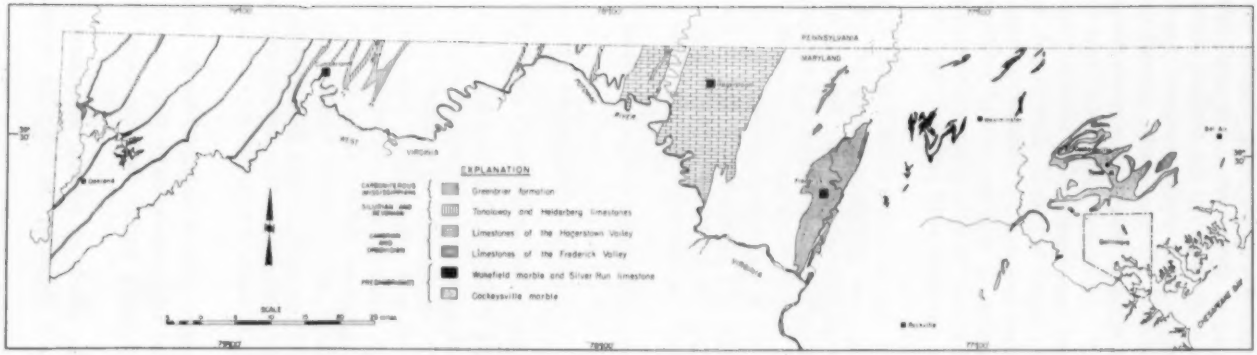

FIG. 1. Geologic map showing rock units which serve as limestone aquifers in Maryland.

of limestone regions. Sinkholes, disappearing streams, and collapsed caverns are almost entirely lacking in these areas. The explanation for this is not clear, but may be related to the fact that the coarsely crystalline marble commonly weathers to calcareous sand, which subsequently is eroded in a manner similar to the more common siliceous sand. Nevertheless, the marble is soluble in ground water acidulated with carbon dioxide, and solutional weather-

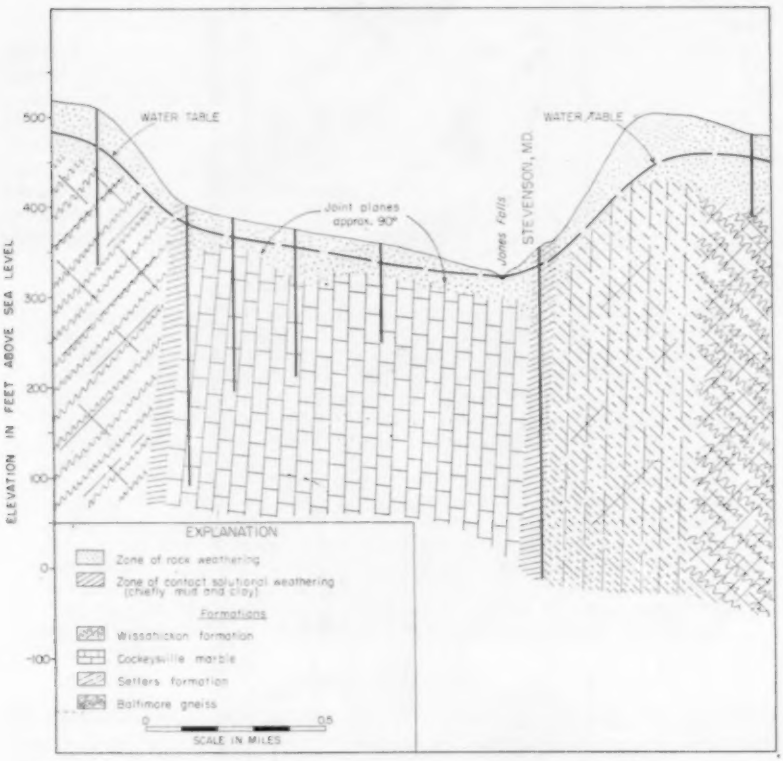

FIG. 2. Geologic section across the Green Spring Valley at Stevenson, Md., showing the depth of rock weathering and the approximate position of the water table in wells.

ing is effective to a degree. This is indicated by the chemical character of the water, which contains substantially more calcium and magnesium bicarbonate and thus is considerably harder than the water from nearby silicate rocks.

The magnitude of the effects of solution of calcareous rocks by ground water in Piedmont areas is shown by the following example. A moderately

large limestone spring in the Green Spring Valley north of Baltimore has a discharge estimated to average about 400 gpm. The dissolved-solids content of the ground water from the marble averages 180 ppm. On this basis it is computed that the water from this single spring carries away about 16,000 tons of dissolved rock per century.

Ground water generally occurs in the limestone rocks of the Piedmont under water-table or localized artesian conditions. The source of the water is almost entirely local precipitation, although some water may move distances of a few thousands of feet laterally along the deeper channels or tubes. That most of the water-bearing channels and crevices occur above depths of 200 feet is indirectly supported by the fact that the average depth of nearly 300 wells in the limestones of Maryland is only 147 feet, although solution of limestones to a depth of 1,000 feet has been reported in deep mines in other areas (17).

Large quantities of ground water are in storage in the limestone aquifers, but reliable values for the storage coefficient are not available for the limestones in Maryland. Walker (18, p. 143) reports a storage coefficient of 0.005 (or 0.5 percent) for a limestone of Mississippian age in Kentucky. If this value is applied to the limestones in Maryland, the uppermost 100 feet of these limestones below the zone of saturation would contain about 100 million gallons in storage in each square mile. This figure, of course, must be regarded as only an approximation illustrating the magnitude of the quantities of water involved. Only part of this water can be economically recovered for use by man.

The geology and hydrology of the major limestone aquifers of Maryland follows.

COCKEYSVILLE MARBLE

Character, Distribution, and Thickness

The Precambrian(?) Cockeysville marble is a gray to white crystalline marble of somewhat variable texture, which ranges from fine-grained to coarse-grained or granular. In places it is dolomitic and micaceous. Some beds have a pinkish or bluish hue. The marble is present chiefly in valleys at Cockeysville and Texas in Baltimore County, where it is extensively quarried (Fig. 1). Narrow outcrop bands extend southward into Howard County and northeastward into Harford County. The marble commonly contains bands of phlogopite and muscovite mica. Other accessory minerals are garnet, asbestos, talc, titanite, and pyrite. Based on analyses of three samples, the average chemical composition of the purest phase of the marble is as follows (10, p. 165):

Constituent	Percent (by weight)
CaO	30.0
MgO	20.5
CO_2	44.7
SiO_2	3.3
Al_2O_3	.9
MnO	.4
Total	99.8

The true thickness of the marble is difficult to ascertain because of folding and faulting of the original sedimentary beds from which the formation was derived. The thickness, however, is believed to be 300 to 500 feet.

Characteristically, the marble weathers to a white to yellow calcareous, micaceous sand, which in surface exposures is covered by a tan or brownish silty clay. The thickness of the weathered or decomposed rock varies greatly, ranging, on the basis of the records of 49 wells, from a few feet to approximately 200 feet. The average depth of weathering in Baltimore County is estimated to be 40 to 45 feet. The development of karst topography and the associated solutional conduits does not appear to take place in areas underlain by the Cockeysville marble. However, weathering, both solutional and mechanical, does take place more rapidly in the marble than in the adjacent gneisses and schists. For this reason most of the localities underlain by marble are valleys or basins.

Hydrology

The Cockeysville marble is an important aquifer in the Piedmont of Baltimore County, where it is tapped by several hundred drilled and dug wells. A large number of springs issue from it, the discharge of a few being large enough for the water to have been used and sold commercially. The comparatively high yield of wells in the marble is believed to be due to the relatively high permeability of the calcareous sand present where the rocks are extensively weathered and where the disintegrated rock material has not been removed by erosion.

Details of the character of the weathered marble and of the rock structure in relation to wells are shown by Figure 2, which is a geologic section in the vicinity of Stevenson, Md. Figure 2, based largely on well logs, shows that the extent and depth of weathering in the marble vary widely within horizontal distances of a few hundred feet. A few wells situated near the limestone valley walls have been drilled to depths of nearly 500 feet without encountering hard rock. These wells commonly penetrate clay, mud, and clayey sand throughout much of their depth. Several of them have been unsuccessful and have been abandoned because the water is muddy. The explanation for the muddy condition is not known for certain, but it may be related to the crushing and disintegration of the rock which may have resulted from faulting and slippage, or it may be largely the result of the chemical disintegration of the marble caused by downward-moving slightly acid ground water from the adjacent silicate rocks. Field evidence neither supports nor eliminates the possibility that faulting has occurred along the contact zones.

Particle-size analyses of five samples of the sand weathered from the marble, made in the Denver hydrologic laboratory of the U. S. Geological Survey, indicate that the sand is chiefly medium to coarse. Permeability tests indicate an average of 640 Meinzer's units, or meinzers (gallons per day per square foot). These values are in the range encountered in the sands of the Coastal Plain in the Baltimore area (2, p. 51–52).

Wells.—The most common diameter of wells in the formation is 6 inches. Most wells are constructed by cable-tool rigs, the well casings being driven

to hard rock. The depths of 54 wells range from 50 to 1,800 feet and average 210 feet. The deepest well was drilled to a depth of 1,800 feet in 1910 near Lutherville, Md., for a proposed municipal supply, but it was never used because its yield was inadequate. The yields of wells range from 1 to 80 gpm and average 19 gpm. The specific capacity (yield per foot of drawdown of the water level) averages 1.0 gpm per foot. Based on 54 wells, the yield per foot of well drilled is about 0.1 gpm per foot—that is, wells averaging 190 feet in depth yield an average of 19 gpm. However, as the water enters a limestone well by means of water-filled tubes, crevices and similar openings, its yield is governed chiefly by the frequency and size of these openings, and its yield cannot be predicted in advance of drilling. In dense, tight rock where these openings are lacking, little or no water will be obtained from drilled wells. On the other hand, a single sizable opening may furnish all the water yielded by an above-average well. Generally, the openings disappear at depths of more than a few hundred feet and the yield per foot decreases with increasing depth.

Springs.—Many springs issue from the marble along the valley walls where it is in contact with the crystalline silicate rocks. The existence of these springs is due to favorable local geologic and topographic conditions. Downward moving ground water may be diverted in a horizontal direction along partially clay-filled fractures or joint planes, or along the top of zones of less pervious rock; such water may thus emerge at the surface as springs or seeps. The flow of these springs varies widely throughout the year owing chiefly to variations in precipitation. The rate of discharge may vary by more than several hundred percent throughout a single season.

Meinzer (13, p. 53) proposed a system of spring classification based on the magnitude of the average discharge. The classification of springs under this system is as follows:

Rank of spring	Discharge
1st magnitude	More than 100 cfs
2nd magnitude	10 to 100 cfs
3rd magnitude	449 to 4,490 gpm (1 to 10 cfs)
4th magnitude	100 to 449 gpm
5th magnitude	10 to 100 gpm
6th magnitude	1 to 10 gpm
7th magnitude	¼ to 1 gpm
8th magnitude	Less than ¼ gpm

Only one spring in the Cockeysville marble is known to be of the 4th magnitude (it was flowing 400 gpm in April 1954). Ten inventoried springs are of the 5th magnitude and three are of the 6th. There are, however, many springs of the 6th, 7th, and 8th magnitudes of which no record was obtained, as their flow is too small or intermittent to warrant utilization and observation.

Quality of the Water

Water from the Cockeysville is generally satisfactory for most uses. It is a mildly alkaline hard calcium bicarbonate water. The range in the pH of 7 samples is from 6.0 to 8.1, and only 1 sample was acidic in character. The average solids content of 8 samples is 180 ppm. The average hardness (as

$CaCO_3$) of 8 samples is 134 ppm. The average total iron content of 8 samples is 0.2 ppm. The silica (SiO_2) content of 4 samples averages 12 ppm.

WAKEFIELD MARBLE AND SILVER RUN LIMESTONE

Character, Distribution, and Thickness

The Precambrian(?) Wakefield marble, a formation of the Glenarm series, is a pure white to grayish-white finely crystalline rock composed chiefly of calcite or dolomite and some impurities. It crops out in many narrow, irregular bands in eastern Frederick and western Carroll Counties (Fig. 1). The maximum width of its outcrop belt is 1 mile in the Wakefield Valley a few miles east of New Windsor in Carroll County. In places the marble contains interbedded white quartzite, and where it is near the adjacent volcanic rocks its color may be white, blue, or mottled pink and green. The mottled marble locally has been called "calico" or "copper" marble. The marble is quarried in at least two places and has been used for cement making. Based on 15 samples the average composition is as follows (12):

Constituent	Percent (by weight)
CaO	52.3
MgO	1.4
CO_2	42.8
SiO_2	2.3
Al_2O_3	.9
Fe_2O_3	.2
Total	99.9

The structural relationship of the marble to the overlying and underlying rocks is very complex. Dips are steep and in many places the marble appears to be sandwiched in between segments of the Catoctin metabasalt and in other places it lies adjacent to phyllites and schists (9). The maximum thickness of the formation is not known, but it has been estimated to be at least 150 feet near Union Bridge in Frederick County.

The Silver Run limestone is a blue thin-bedded finely crystalline limestone of Precambrian(?) age. It has partings of blue shale, is folded, and is veined with calcite and quartz. It weathers to a clayey soil. It is believed to be approximately equivalent in age to the Wakefield marble. The Silver Run limestone crops out in irregular north-trending bands between Union Bridge and Westminster, Md. The total area of its outcrop is estimated to be not more than 3 square miles. No data are available regarding its thickness, but it probably is not more than a few hundred feet thick.

Hydrology

With regard to the occurrence of ground water, the Wakefield marble and the Silver Run limestone are treated as one unit in this paper. The formations are important sources of water for farm, domestic, and small industrial supplies. The water is derived largely from local precipitation and occurs under water-table or localized artesian conditions. In some wells the water is re-

ported to have been encountered in dark brown "sand" or a bed of "coarse gravel." It is likely that sand is disintegrated crystalline marble and that the gravel is disintegrated quartz veins or thin beds of siliceous limestone.

Many of the common topographic features of limestone regions are absent in localities underlain by the Wakefield marble and Silver Run limestone. Sinkholes and disappearing streams do not appear to be present, although Davies (7, p. 29) reports the existence of a small cave in the Wakefield near Westminster, which may be of solutional origin.

Wells.—On the basis of well yields, the Wakefield marble and Silver Run limestone are the best limestone aquifers in Maryland. The average depth of wells is somewhat less than in the Cockeysville marble. Forty-one wells range in depth from 8.7 to 575 feet and average 132 feet.

The yields of wells range from 1½ to 575 gpm and average 90 gpm. Meyer (14) reports that in 1948 two wells in a narrow valley at the Koontz Creamery Co. at Westminster yielded 575 and 540 gpm, respectively. The yield of 575 gpm was obtained from a 10-inch well penetrating cavernous marble at a depth of approximately 160 feet. Owing to flowing mud entering the hole, the well was subsequently deepened to 850 feet, the marble cased off, and a later test at the greater depth yielded only 7 gpm. The well was subsequently abandoned. The specific capacities of 17 wells range from 0.1 to 22 gpm per foot and average 4.8 gpm per foot. Based on the records of 30 wells, the average yield per foot of well drilled is 0.7 gpm. This is the highest value for wells in any limestone aquifer in Maryland.

Springs.—Springs are fairly common in the Wakefield marble but records of five of them indicate that their flow is not large. Three of the springs are of the 5th magnitude (10–100 gpm) and two are of the 6th magnitude (1–10 gpm).

Quality of the Water

The chemical character of the water is generally satisfactory. It is a hard, slightly alkaline calcium bicarbonate water. The range in pH of 4 samples was from 7.4 to 7.7. The dissolved solids content of two samples was 170 and 194 ppm. The average hardness of 4 samples was 123 ppm. Average total iron content of the same samples was 0.1 ppm.

LIMESTONES OF THE FREDERICK VALLEY

Character, Distribution, and Thickness

Three calcareous formations are present in the Frederick valley; they are the Tomstown dolomite and the Frederick limestone of Cambrian age, and the Grove limestone of Ordovician age. The Tomstown is a white crystalline dolomite which commonly weathers to a red clay. In places it contains sericite and carbonaceous partings. Its thickness in the Frederick valley is about 200 feet.

The Frederick limestone is commonly dark blue in color and consists of thin coarsely crystalline layers separated by dark thin argillaceous partings. A gray to black shale, which occurs near the base of the Frederick, crops out

along the east side of the valley. The Frederick limestone commonly weathers along joints and bedding planes into tabular blocks or boulders. The end product of continued weathering is a brown to reddish silty clay which in places may be as much as 35 feet thick. The total thickness of the Frederick limestone is about 500 feet.

The Grove limestone consists, in the upper part, of thick beds of high-calcium limestone. Near the base, beds of massive dolomite and highly quartzitic limestone predominate. In places the silica content of the basal beds is sufficiently high that the rock weathers to a porous sandstone which, where disintegrated, is used for building sand. Several quarries are located in the Grove limestone where it is pure and high in $CaCO_3$. An analysis of a composite sample from a quarry at LeGore is as follows (12, p. 385):

Constituent	Percent (by weight)
CaO	49.4
MgO	3.0
CO_2	41.8
SiO_2	4.6
Al_2O_3 and Fe_2O_3	1.1
Total	99.9

Based on a section measured in the LeGore quarry, the thickness of the Grove limestone is about 600 feet. Thus, the total thickness of the three calcareous formations in the Frederick valley is about 1,300 feet.

Structurally, the Frederick valley may be described as a complex synclinorium in which faulting and repetition of individual beds are common. The strike of the beds is commonly north-northeast, but changes in places to north or northwest. In the vicinity of Frederick, the dip of the strata ranges from 5 to 50 degrees from the horizontal (9). It is likely that the downward movement of ground water has been greatly facilitated locally where the rocks dip steeply. Solutional openings might be expected at considerable depth in such places.

Hydrology

Ground water is an important resource in the Frederick valley, as nearly all the residents not served by the public supplies of Frederick and other towns with surface sources depend on ground water. In addition, the public supplies of Thurmont, Walkersville, and Woodsboro are obtained chiefly from limestone wells and springs.

Wells.—The depths of 153 wells in the valley range from 15 to 1,209 feet and average about 123 feet. The yields of 107 wells, both domestic and industrial, range from 2 to 275 and average 27 gpm. Specific capacities of 58 of these wells range from 0.1 to 43 and average 3.5 gpm per foot of drawdown. The yield per foot of 105 wells of all types is 0.25 gpm, which is near the value obtained for the 9 best wells. Meyer (14) reports that the average yield of 9 large-capacity wells is 125 gpm. The average depth of these wells is 358 feet and the yield per foot of well drilled is 0.35 gpm.

Springs.—Comparatively few spring records are available from the Frederick valley. One spring issues from a cavern in the Grove limestone south-

west of Woodsboro. On the basis of an estimated flow of 100 to 200 gpm in October 1955, it is a spring of the 4th magnitude. Undoubtedly many other springs exist for which records are not available.

Quality of the Water

Ground water from the limestones of the Frederick valley is a very hard calcium bicarbonate water, 13 samples having an average hardness of 234 ppm. The range in pH of the 13 samples is from 7.4 to 8.0. Dissolved solids are low, averaging only 274 ppm. for 7 samples. The iron content in 13 samples averages 0.1 ppm., which is less than the suggested limit of 0.3 ppm set by the U. S. Public Health Service for potable water used on interstate carriers. The silica content (SiO_2) averages 8.8 ppm for 7 samples. This amount is below the recommended maximum limit of silica in water used for icemaking and boiler-feed water except where extremely high boiler pressures are required (4, p. 353).

LIMESTONES OF THE HAGERSTOWN VALLEY

Character, Distribution, and Thickness

Seven limestone formations are exposed in the Hagerstown valley in west-central Maryland. These are the Tomstown, Waynesboro, Elbrook, and Conococheague formations, of Cambrian age, and the Beekmantown, Stones River, and Chambersburg limestones, of Ordovician age. The oldest unit, the Tomstown dolomite, has been described previously. In Pennsylvania its thickness has been reported as 1,000 feet. It is overlain by the Waynesboro formation which consists of two units, a lower unit of dolomite and shale and an upper unit of shale and sandstone. Northward in Pennsylvania there are reportedly conglomerate beds in the lower unit. Commonly the upper shales and sandstones are red, green, and yellow, and are ripplemarked and cross-bedded. The gray limestones at the base are siliceous and usually weather to porous sandstones. The formation is at least 600 feet thick in the Hagerstown valley near Chewsville. Its thickness in Franklin County, Pennsylvania, is reported not to be in excess of 1,000 feet (8, p. 50).

The Elbrook limestone, which overlies the Waynesboro formation, is a thick series of gray to light-blue shaly limestones and calcareous shales. Massive beds of limestone and dolomite, siliceous limestone, and siliceous oölite are present in the middle of the formation. The siliceous limestones commonly weather to porous sandstone. In Maryland complete sections are not exposed, but Butts (3, p. 74) reports that south of the Potomac River the unit is 1,400 to 2,000 feet thick.

The Conococheague, overlying the Elbrook, consists mostly of dark blue impure limestones, bands of shale and conglomerate, beds of oölite, and, near the base, layers of sandstone. Near the top, pink marble and light-colored limestone occur. The formation is approximately 1,800 to 2,000 feet thick.

The Beekmantown limestone consists of purer and more massive beds of limestone than the Conococheague on which it lies. The basal beds consist

of minutely laminated pink and white marble, siliceous banded beds, and thin edgewise conglomerates. The thickness of the formation in Maryland is reported by Cloos (5, p. 59) to be about 3,000 feet.

The Stones River limestone, St. Paul group of Neuman (15) consists of a series of granular impure limestones, which locally contain chert. This series is overlain by fine-grained dove-gray and dark limestone. Hall (8, p. 53) reports the thickness of the unit in the nearby Chambersburg area, Pennsylvania, to be 650 to 1,050 feet.

The Chambersburg limestone, the youngest unit in the Hagerstown valley, consists of 100 to 750 feet or dark gray medium-grained thin-bedded agrillaceous limestone. When weathered it may be recognized in the field by its cobbly appearance.

It is estimated that the total thickness of limestones in the valley is between 8,500 and 10,400 feet. Structurally, the limestones occupy a complex synclinorium which attains a width of nearly 20 miles near the Pennsylvania line. The strata strike north-northeast and their dips range in most places from 5° to 80°. In places, however, the beds are overturned. The Martinsburg shale, essentially a noncalcareous unit, is exposed along a 2½- to 4-mile wide belt in the western part of the valley. The shale may be regarded as infolded in a sequence of limestone beds. Joints and openings along bedding planes are common in the limestones. It is likely that the local occurrence of ground water is governed chiefly by nature, distribution, and vertical extent of the rock joints. Solutional features may be expected to occur where downward-moving water enlarges the initial fractures, and successful wells may be expected where water-saturated openings are penetrated by the drill.

Hydrology

Most rural and domestic water supplies in the valley are obtained from wells or springs. Most of the water in the upper few tens of feet of limestone probably occurs under water-table conditions. The major source of the water in the valley is local precipitation. However, the ground water is in continual movement from the interstream recharge areas to the surface streams where it is carried away from the valley. In general, it is believed that little ground water is stored in the rocks below depths of 300 to 400 feet.

Wells.—Thirty-four wells in the area range in depth from 25 to 920 feet and average 202 feet. The yield of 31 wells ranges from 0.1 to 200 gpm and averages 48 gpm, but these wells include high-capacity industrial wells in and near Hagerstown. Cloos (5, p. 184) reports the average yield of 260 wells, chiefly domestic, to be about 8 to 10 gpm. The specific capacity of 20 of the 31 wells ranges from < 0.1 to 21 and averages 3.8 gpm per foot. Based on the yields of the 31 wells penetrating 5,900 feet of rock, the yield per foot of well drilled is 0.25 gpm. This is the same value obtained from the records of 105 wells penetrating 18,700 feet of limestone in the Frederick valley to the east.

Springs.—Many large springs occur in the Hagerstown valley, and the communities of Boonsboro, Clear Spring, and Smithsburg are supplied by

springs. A spring apparently issuing from the Tomstown dolomite supplies an orphanage and a rest home near Sanmar. Most of the springs in the valley are of the 5th magnitude (10 to 100 gpm) or less, but an exceptionally large spring issues from the Beekmantown limestone at Big Spring. This spring is estimated to be of the 3rd magnitude, although to date its flow has not been measured.

Quality of the Water

On the basis of only four analyses, ground water from the Hagerstown valley appears to be a very hard calcium bicarbonate water, averaging 315 ppm in hardness. The range in pH of four samples is from 7.2 to 7.7. The average dissolved solids content is 392 ppm and the average iron content is 0.4 ppm. The silica (SiO_2) content averages 17 ppm in four samples. Softening of the water would be required for most uses.

TONOLOWAY AND HELDERBERG LIMESTONES

Character, Age, and Thickness

The Tonoloway limestone is a fine-grained dark gray limestone that weathers to light gray or buff. The conspicuous thin laminations are shale partings. In its eastern exposures the formation consists of limestone and shale, separated by a thin sandstone bed about 120 feet above the base of the formation. The Tonoloway is Silurian in age. Near Hancock it is about 375 feet thick. It thickens to the west and in some Garrett County wells is almost 700 feet thick.

The Helderberg is a blue or gray limestone with chert, sandstone, and numerous fossils in some beds. The Helderberg limestone is Early Devonian in age and its average thickness is about 400 feet. However, its thickness varies considerably from east to west.

Distribution

For the purpose of this report the Tonoloway and Helderberg are considered to be one unit and are so indicated on the accompanying geologic map. This has been done in some areas in Pennsylvania as well (11, p. 95).

In Maryland these formations crop out only in Allegany and Washington Counties. Despite their calcareous nature they occur fairly high on parallel ridges that characterize the topography in most of Allegany and western Washington Counties, where they usually crop out in long, narrow bands. In Allegany County they occur on Fort Hill, Haystack and Wills Mountains, Shriver Ridge, and Evitts, Martin, and Warrior Mountains. In Washington County the limestones crop out on Tonoloway Ridge, Roundtop, Cove Ridge, Elbow Ridge, at the south end of the Bear Pond Mountains, and on Fairview Mountain.

Hydrology

At the present time, almost no information is available on the water-bearing properties of the Tonoloway and Helderberg limestones in Maryland. Field work in these areas is to be started in the near future.

Wells.—Twenty-one wells in the Tonoloway and Helderberg limestones in south-central Pennsylvania range in depth from 27 to 408 feet and average 114 feet. Yields of 19 of these wells range from about 4 to 200 gpm and average 22. Excluding two wells having yields of 200 and 125 gpm, these wells have an average yield of 5 gpm. Specific capacities are available for only 2 wells: one is less than 0.1 and the other is 1.4 gmp per foot of drawdown. The yield per foot of well drilled for 19 wells averages 0.2 gpm.

Springs.—The discharge of 8 springs from these limestones in south-central Pennsylvania ranges from 30 to 2,000 gpm and averages about 140. Four of these are 5th-magnitude springs, three are of the 4th magnitude, and one is of the 3rd magnitude.

Quality of the Water

Inasmuch as the Tonoloway and Helderberg are both limestones, one would expect the water from these formations to be hard. Analyses show that this is true, but there is, however, a wide range in mineral content, no doubt due to variations in lithology. The following analyses from springs in Pennsylvania illustrate this (11, p. 139 and 151):

	Spring 509 (Tonoloway ls.)	Spring 992 (Helderberg ls.)
Ca	34 ppm	558 ppm
Mg	—	146 ppm
Na + K	3 ppm	5 ppm
HCO_3	164 ppm	183 ppm
SO_4	13 ppm	1,764 ppm
Cl	1.0 ppm	6.0 ppm
NO_3	1.0 ppm	.20 ppm
TDS	158 ppm	2,565 ppm
$CaCO_3$	144 ppm	1,994 ppm

GREENBRIER FORMATION

Character, Distribution, and Thickness

The Greenbrier formation consists chiefly of fossiliferous calcareous shale and sandstone and some argillaceous and arneaceous limestone. The shales are usually red, brown, green, or mottled red and green. The lower part of the Greenbrier has been named the Loyalhanna member; this is generally a medium- to fine-grained red to gray limestone with quartz sand; crossbedding is a prominent feature in some places.

The Greenbrier is underlain by the Pocono formation, the contact being placed, according to Amsden (1, p. 26–27) "where the non-calcareous sandstones of the Pocono are overlain by the calcareous shales or sandy limestones of the Greenbrier formation." It is overlain by the generally noncalcareous, unfossiliferous shale, sandstone, and siltstone of the Mauch Chunk formation.

Although there has been some dispute over the exact age of the Greenbrier formation, it is considered to be Mississippian. The unit crops out in only two counties in Maryland, Garrett and Allegany. In Garrett County it crops out on the flanks of the Accident and Deer Park anticlines in exposures rarely

more than 0.2 or 0.3 mile wide. The narrowness of the outcrop belts is due to the fairly steep dips in almost all localities except at the south end of the Accident anticline, where the exposure is about a mile wide, and to the fact that the formation is only 200 to 300 feet thick. In Allegany County the Greenbrier crops out only along Dans Mountain, on the east side of the Allegheny Front. Here the outcrop is rarely wider than about a tenth of a mile.

The Greenbrier and the overlying Mauch Chunk have little resistance to erosion and therefore neither is well exposed. The best exposures are in quarries where limestone is being removed, and at Savage River Dam.

Hydrology

Because of its small areal extent and its occurrence in generally sparsely populated areas, the Greenbrier is not an important aquifer, and relatively few wells have been drilled to it. In Garrett County 16 wells ending in the formation range in depth from 45 to 164 feet and average 82 feet. The yields of 13 wells ending in the Greenbrier range from 1 to 20 and average 13 gpm. The specific capacities of 6 wells range from 0.2 to 3.6 gpm per foot and average 1.3. For 13 wells the yield per foot of well drilled averages 0.2 gpm.

Springs.—Most springs inventoried in the Greenbrier formation are contact springs at the contact of the Greenbrier with either the underlying Pocono or the overlying Mauch Chunk formation. The discharge of 7 springs ranges from 0.2 to 50 and averages 12.5 gpm. One of these is of the 7th magnitude, four are of the 6th magnitude, and two are of the 5th magnitude. Boiling Spring, in Garrett County, has the largest flow, 50 gpm. It is owned by the Baltimore and Ohio Railroad and its water is used in dining cars.

Quality of the Water

Chemical analyses of water from the Greenbrier formation are not available. Inasmuch as there is considerable variation in the lithology of the Greenbrier formation, there is doubtless considerable variation in the chemical quality of the ground water.

U. S. Geological Survey,
Baltimore, Maryland,
Dec. 7, 1957

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PORPHYRITIZATION IN DESTOR AND DUPARQUET TOWNSHIPS, ATIBI WEST COUNTY, QUEBEC, CANADA[1]

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ABSTRACT

In Destor and Duparquet townships, Precambrian quartz-feldspar porphyry intrusives have been injected into Keewatin-type lavas and pyroclastics, Timiskaming-type conglomerate and arkose and post-Keewatin type, pre-Timiskaming type basic intrusives.

Surrounding the quartz-feldspar porphyry intrusives are halos containing numerous well developed quartz crystals. These are present in all types of rock with which the quartz-feldspar porphyry is in contact.

The presence of quartz crystals in conglomerate boulders adjacent to the porphyry and the presence of quartz-feldspar porphyry fragments and conglomerate boulders along brecciated contacts between conglomerate and quartz-feldspar porphyry had led previous workers to the erroneous conclusion that the Timiskaming-type sediments are younger than the quartz-feldspar porphyry.

A study of the above features as well as the halos of quartz crystals

[1] Presented with the permission of the Deputy Minister, Department of Mines, Quebec.

developed in other rock types adjacent to the quartz-feldspar porphyry indicates that these phenomena are due to porphyritization.

Porphyritization is the development of a porphyritic and porphyroblastic texture in a previously formed rock by partial magmatic replacement and by metastomatism.

The process accompanying porphyritization has been divided into five stages:

(1) The injection of quartz-feldspar porphyry.
(2) Permeation of the more porous rock types at the contact by a tenuous phase of the quartz-feldspar porphyry liquid fraction.
(3) Brecciation and fracturing.
(4) Injection of a highly siliceous fraction of quartz-feldspar porphyry into zones of fracturing and brecciation.
(5) Injection of the final liquid residue of the cooled quartz-feldspar porphyry into zones of fracturing and brecciation.

Gold deposition took place in these zones, at this stage, and is genetically related to the quartz-feldspar porphyry.

INTRODUCTION

PORPHYRITIZATION is a phenomenon of partial magmatic replacement or metasomatism associated with certain quartz-feldspar porphyry intrusives in Destor and Duparquet townships. Field observations and material for laboratory study were obtained during the summers of 1944 and 1945 while the writer was carrying out detailed geological mapping for the Quebec Department of Mines. The laboratory work was carried out at the University of Toronto. The area mapped is comprised of some 25 square miles situated 20 miles north of Noranda, Figure 1, and 335 miles north of Toronto, Ontario.

Geological work in the region has been carried on from time to time since 1872 (see references). As a result of this work, the age of a series of porphyry intrusives relative to the Duparquet-Timiskaming type sedimentary formation has been the subject of much debate. The preponderance of detailed field evidence leaves little doubt that the porphyries are younger than the Duparquet sidements; yet there are certain anomalous relationships, such as the presence of porphyry "boulders" in the conglomerate adjacent to the porphyry and the presence of local areas of intermingled porphyry fragments and conglomerate boulders along zones of faulting called "porphyry conglomerates." Because of this, a careful study was made of the contacts of the Duparquet sediments with the porphyries. It was found that these anomalous relationships were confined to the vicinity of one type of porphyry, namely, quartz-feldspar porphyry. Further, at the contacts of this porphyry; the "porphyry conglomerate," the sediments and the porphyry boulders they contain exhibit features that cannot be explained as being derived from an older porphyry terrain. Additional study has also found these features to be present in all other rock types in close proximity to the quartz-feldspar porphyry.

As a result of these studies, the theory of porphyritization has been advanced as an explanation of the relationships between quartz-feldspar porphyry and conglomerate. This theory also offers an explanation for the association of gold with the porphyries in the general district.

GENERAL GEOLOGY

The consolidated formations are all Precambrian in age (Fig. 1). The oldest is a Keewatin-type group that has been subdivided into the Older Volcanics, the Cléricy Sediments and the Younger Volcanics.

The Older Volcanics consist predominantly of a lower formation of andesitic and trachytic lavas with some rhyolite and basalt flows succeeded conformably by an upper formation consisting predominantly of basaltic flows.

Following a period of gentle folding and erosion, the Cléricy Sediments, comprised of slate and graywacke, were deposited on the Older Volcanics.

The Younger Volcanics and related intrusives were laid down conformably on the Cléricy Sediments. These consist of a lower formation composed predominantly of basaltic flows with some andesite and trachyte, and an upper formation consisting predominantly of andesitic and trachytic flows with some basalt, agglomerate, and tuff.

In the eastern part of the map-area, two tabular bodies of gabbro, quartz gabbro and pyroxenite were intruded into the Keewatin-type formations. These are known as the Pre-Duparquet intrusives.

Following this period of intrusion, there was a period of folding and erosion. The Duparquet sediments consisting of conglomerate and arkose were then laid down and are preserved as a belt that extends centrally across the area. After deposition of the sediments, there was folding resulting in the formation of the Lépine lake syncline. Subsequently, a period of gentle folding took place. Regional evidence suggests that this period of folding is related to the forces that produced the Destor-Porcupine fault and its subsidiary zones. This fault, trending slightly north of west, comprises the north or lower contact of the Cléricy sediments. Where these sediments have pinched out, it forms the line of separation between the Older Volcanics to the north of the fault and the Younger Volcanics that lie to the south.

Towards the close of the faulting bodies of feldspar porphyry and quartz-feldspar porphyry, granite and aplite were intruded as well as small dikes of lamprophyre. These occur along and adjacent to the various zones of faulting. They are known as the post-Duparquet intrusives.

The area was then subject to a second period of widespread faulting although on a smaller scale. Finally, many small dikes of diabase, gabbro, diorite and some pyroxenite were introduced. Their orientation was apparently controlled by the fracturing developed during the second period of faulting.

Since this paper is mainly concerned with the relationships between the Duparquet sediments, and certain members of the post-Duparquet intrusives, these formations will be described in greater detail than that provided in the preceding paragraphs.

Duparquet Sediments.—The Duparquet sediments outcrop centrally across the map-area. Conglomerate comprises the greater part of these sediments. Typically, it is poorly sorted but local exposures of pebbles of a fairly uniform size were observed. Rounded to slightly elongated pebbles and boulders up to 3 feet in diameter make up 85 to 95 percent of the rock. The matrix con-

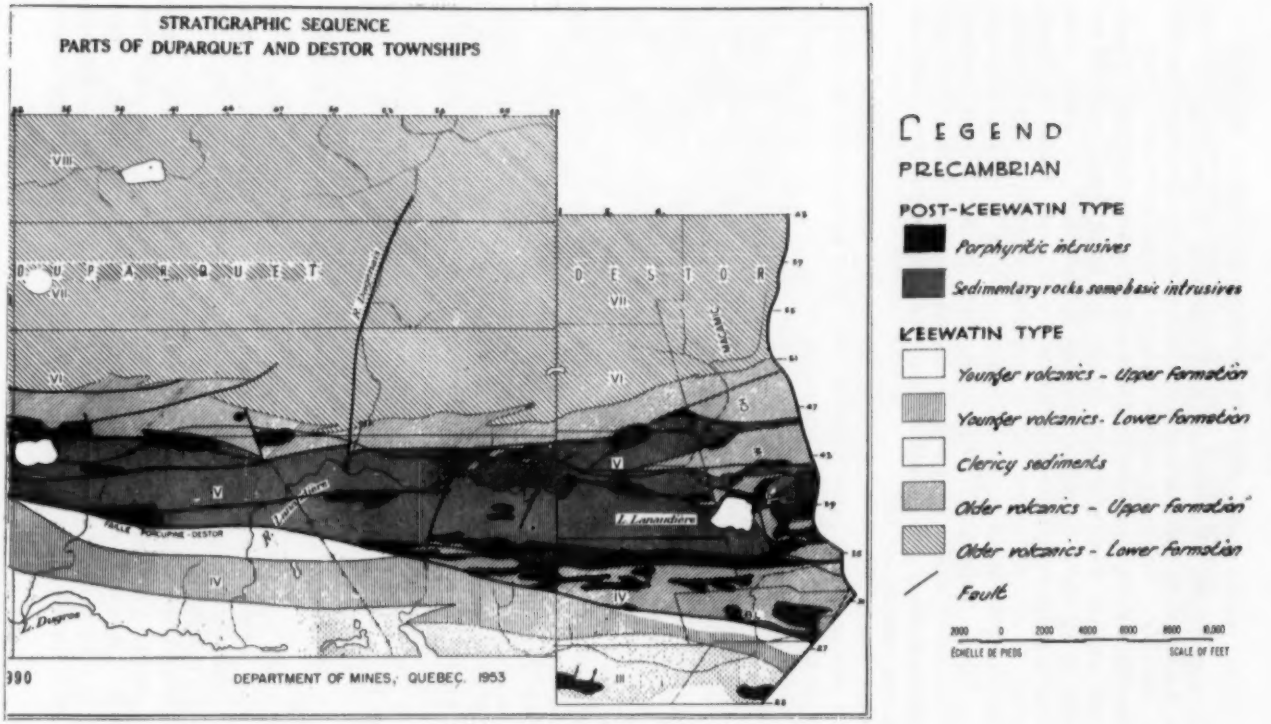

STRATIGRAPHIC SEQUENCE
PARTS OF DUPARQUET AND DESTOR TOWNSHIPS
DUPARQUET
DESTOR
VIII
VII
VI
V
IV
III
L. Lanaudière
L. Dugras
DEPARTMENT OF MINES, QUEBEC, 1953
LEGEND
PRECAMBRIAN
POST-KEEWATIN TYPE
Porphyritic intrusives
Sedimentary rocks some basic intrusives
KEEWATIN TYPE
Younger volcanics - Upper formation
Younger volcanics - Lower formation
Clericy sediments
Older volcanics - Upper formation
Older volcanics - Lower formation
Fault
ÉCHELLE DE PIEDS
SCALE OF FEET

FIG. 1.

sists of arkosic material. Boulders in the conglomerate consist of graywacke, argillite, gabbro, diabase, gray-granite, chert, basalt, andesite, jasper, rhyolite, quartz-feldspar porphyry and amphibolite.

Arkose forms a minor constituent of the Duparquet sediments. A few slaty bands are present with the arkose, but these are local in distribution. Arkose horizons are found mainly along the trace of the synclinal axis which passes through the sediments. Bedding within individual horizons is variable and local. It may be lens-like, or it may consist of regular bands up to several feet in width. Scour channelling, cross bedding and grain grading are present in places.

Post-Duparquet Intrusives.—The post-Duparquet intrusives are distributed, for the most part, along a zone, with a maximum width of about 1½ miles, crossing centrally through the area in a direction slightly south of east. This zone follows the zone of fracturing and faulting that comprises the Destor-Porcupine fault. A few isolated bodies lie outside this zone both to the north and south.

The porphyry intrusives occur as lens-like bodies associated with shear zones and at the intersections of such zones. In detail, these bodies are irregular in outline, pinching and swelling in both horizontal and vertical extent. Where numerous, more or less parallel, shears form a zone several hundred feet wide, the porphyry may be intruded in the form of many sill-like lenses with local cross-cutting relations. These intrusives contain inclusions of conglomerate, of members of the pre-Duparquet intrusives and of sheared volcanics.

Included in the post-Duparquet intrusives are micrographic feldspar porphyry, feldspar porphyry, quartz-feldspar porphyry, aplite and albite granite. There are two ages of feldspar porphyry. One, older than the quartz-feldspar porphyry, comprised predominantly of albite phenocrysts in an albitic matrix, and the other, younger than the quartz-feldspar porphyry, containing phenocrysts of albite and microcline in a matrix of albite and quartz. The aplite is a phase of the older feldspar porphyry. The age of the micrographic feldspar porphyry and the albite granite with respect to the other members of the group is not known.

Quartz-Feldspar Porphyry.—The quartz-feldspar porphyry is of particular interest because of its bearing on the phenomena of porphyritization.

Typical exposures of quartz-feldspar porphyry weather various shades of gray to pink and are massive in appearance. On the fresh surface, it is light greenish-gray, sea green or pink. It has a fine-grained compact texture with well developed squarish quartz phenocrysts up to ⅝ inch in diameter and well developed laths of feldspar from 1/16 to ⅛ inch in length. Generally the feldspar phenocrysts make up about 40 percent of the rock and along with the quartz phenocrysts, impart an apparent coarse-grained texture.

In thin section, both the quartz and feldspar phenocrysts are well developed. The feldspar phenocrysts are of albite and vary in size from 0.25 to 4 mm in length. They are speckled and seamed with white mica. Quartz phenocrysts are present, but not abundant, generally forming only 1 percent of the

Note: In Figure 1 "Younger Volcanics—Upper Formation" is No. III, extending below IV. Diagram not shown in legend.

rock. They may be embayed, with knife-edge contacts against the matrix, or they may have lace-like borders. They are larger than the feldspar phenocrysts, averaging 4 to 5 mm in diameter.

The groundmass consists of a intimately associated aggregate of quartz and white mica. The average grain size of this aggregate is 0.02 mm or less in diameter. Chlorite plates are characteristic of the porphyry and occur as well developed books from ⅛ to ¼ inch across. Apatite is also a characteristic accessory mineral, commonly forming from 2 to 5 percent of the section. One crystal was noted 0.5 mm in length, but generally they are 0.2 mm or less. The apatite is present in the groundmass, in the chlorite plates and in some of the more highly altered feldspar phenocrysts. Leucoxene altered from ilmenite, pyrite, calcite and magnetite form the remainder of the matrix.

Below is listed the chemical analysis of a specimen of the typical quartz-feldspar porphyry.[2] With these results and the known mineralogical composition of the specimen as a basis, the mode of the rock was calculated.

		Mode	
SiO_2	69.37%	Quartz	28.08%
Al_2O_3	17.05	Albite	43.24
CaO	1.04	Anorthite	0.27
MgO	1.20	Sericite	10.64
Fe_2O_3	0.59	Paragonite	6.89
FeO	1.09	Chlorite	3.18
K_2O	1.93	Apatite	2.60
Na_2O	5.53	Calcite	1.40
TiO_2	0.28	Ilmenite	0.45
MnO	0.02	Magnetite	0.69
P_2O_5	0.12	Pyrite	2.14
H_2O+	1.28		
H_2O-	0.05	Total	103.81
CO_2	0.64		
Total	100.19		

An average of 3 Rosiwal analyses of the phenocrysts in the specimen gave the following results: albite as phenocrysts, 40.58%; quartz as phenocrysts, 0.93%. These results checked with the proportion of quartz and feldspar occurring as phenocrysts in other specimens of typical quartz-feldspar porphyry. Thus, approximately 94 percent of the total feldspar content and only 3 percent of the total quartz content occur as phenocrysts in typical quartz-feldspar porphyry.

DEVELOPMENT OF THE THEORY OF PORPHYRITIZATION

Surrounding bodies of quartz-feldspar porphyry, there are halos containing numerous well developed quartz crystals. The halos do not generally extend beyond the quartz-feldspar porphyry more than 80 feet, but in rare cases, they may attain a width of 200 feet. The intensity of the development of the quartz crystals decreases proportionately as the distance from the porphyry increases.

[2] Locality: 800 feet south of the Beattie road and 200 feet west of the Destor-Duparquet township line. Analysis by the Quebec Dept. Mines.

These halos have been recognized in the Duparquet sediments, quartz crystals being developed in both the matrix and boulders of the conglomerate; they have also been observed in breccia zones along the edges of the porphyry bodies, in older feldspar porphyry, in gabbro, in chert, in basalt, in trachyte and agglomerate. The most detailed study was carried out along the contacts of quartz-feldspar porphyry with the conglomeritic and non-conglomeritic members of the Duparquet sediments and along the breccia zones.

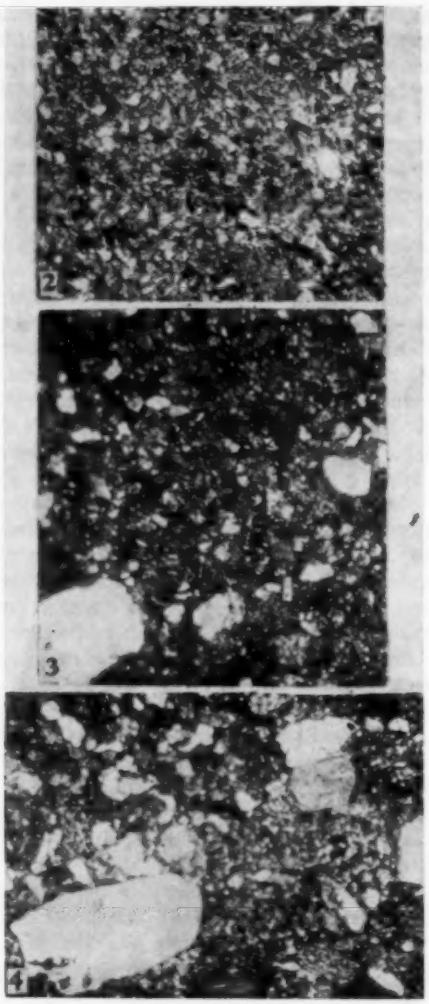
Contact Relations of Quartz-Feldspar Porphyry with Duparquet Sediments.—A typical example of quartz-feldspar porphyry in contact with the Duparquet sediments lies about 1,600 feet west of the Destor-Duparquet township line immediately south of the Beattie road.

The arkose is well bedded. Several of the beds contain numerous stout quartz crystals that project from the weathered surface. The contact between the beds with the quartz crystals and those without is not well defined. A specimen from one of the beds containing quartz crystals was examined in thin section and found to be a normal quartz-feldspar porphyry with no trace of a clastic texture. The thin section was compared with a thin section from a body of massive porphyry 50 feet to the southwest and the two were found to be identical. Thus, it is considered that these particular beds represent sill-like bodies of quartz-feldspar porphyry intruded into the arkose.

Several specimens from the contact between a fine-grained gray-weathering bed of arkose and one of the porphyry sills from the above locality was examined in thin section. The fine-grained arkose consists essentially of quartz, feldspar and sericite. The quartz and feldspar grains have angular outlines and average 0.1 mm in diameter. Also present are a few larger grains averaging 0.5 to 0.6 mm in diameter. The rock is similar in texture and composition to a normal sediment.

As the porphyry side of the section is approached, lens-like patches appear, consisting of a very fine-grained aggregate of quartz, sericite and feldspar, similar to the groundmass of the quartz-feldspar porphyry. Some of the lenses contain well-developed quartz phenocrysts in some cases with embayed outlines, but no feldspar phenocrysts were observed. Occupying the area between the lenses are angular fragments of highly sericitized feldspar, a few angular fragments of quartz and some white mica. Closer to the porphyry the lenses increase in number, the amount of white mica increases and that of detrital quartz decreases. The feldspar fragments are here completely altered to white mica. The detrital quartz and altered feldspar fragments are surrounded by a coating of white mica. Still closer to the porphyry, the lenses become so numerous that they are separated only by selvages of white mica, and quartz and feldspar fragments are no longer present. In the lenses, inclusions of schlieren-like patches of white mica resembling the selvages are not uncommon. Finally, individual lenses of porphyry groundmass can no longer be distinguished and feldspar phenocrysts appear for the first time. The rock is now a normal quartz-feldspar porphyry with a few schlieren-like patches of white mica, which die out into the porphyry (Figs. 2–4).

The above description demonstrates the gradational nature of the contact of a sill-like body of quartz-feldspar porphyry intruded into an arkosic sedi-



Photimicrographs illustrating gradational contact between porphyry and graywacke. Figures 2, 3, and 4 from the same thin section passing from the sediments to porphyry.

FIG. 2. Normal sediment, note size and shape of quartz grains. X-nicols, × 12.9.

FIG. 3. Normal sediment at the top of the photograph. Towards the bottom, "phenocrysts" of quartz appear. These are considerably larger than the detrital quartz of the sediments and have not the angular outline. Towards the bottom of the picture may be seen closely packed lenses of typical fine grained porphyry groundmass. X-nicols, × 13.6.

FIG. 4. Quartz feldspar porphyry. Note the numerous angular quartz fragments included in the groundmass. X-nicols, × 13.6.

ment. In this particular locality the contact zone varies from several inches to 10 feet wide.

Contacts showing identical features to the above, both on the outcrop and in thin section, may be seen in the following localities: Three-quarters of a mile south of the Beattie road and 3,600 feet west of the Destor-Duparquet township line; in lot 39, west range, Macamic road, just west of the range line.

Effect of Porphyry on Boulders in Duparquet Conglomerate.—Adjacent to quartz-feldspar porphyry, many boulders in the conglomerate contain quartz eyes irrespective of their composition. Several specimens of boulders showing these features were examined in thin section.

A pebble of sedimentary origin contained in the conglomerate was examined in thin section. It consists of 60 percent quartz as grains, many of which are angular, 15 percent saussuritized and recrystallized feldspar too altered for identification, 10 percent oligoclase, 10 percent epidote as grains scattered throughout the section, 2 percent chlorite and 2 percent magnetite and pyrite, as grains averaging 0.6 to 1 mm in diameter. This rock has the texture of a sediment that has been somewhat recrystallized. The specimen is cut by quartz filaments that contain minute spherules of quartz. Scattered throughout the section are "phenocrysts" of quartz. The quartz veinlets commonly join in with the "phenocrysts" and where they do, the "phenocrysts" are enlarged by growths of quartz around their margins. These have the same crystallographic orientation as the "phenocrysts."

A specimen of a boulder of tuff in the conglomerate is seamed with long, narrow, lens-like stringers of fine grained quartz. These are locally enlarged to form elongated quartz crystals with fuzzy indefinite lace-like borders.

A boulder similar to a rhyolite formation occurring along the north border of range VIII in Duparquet township contains phenocrysts of albite clouded with saussurite, some of which are mildly fractured. The quartz phenocrysts are of two types. One has sharp, clear-cut borders and is stout and squarish in outline. The other is more euhedral in outline and has delicate lace-like borders suggesting growth by replacement. The groundmass is composed of quartz, feldspar and sericite. The quartz is very fine grained, less than 0.01 mm in diameter and generally has a stringy or wandering nature with delicate lace-like borders similar to that surrounding the phenocrysts of the second type described above.

Contact Relations with Porphyry-Conglomerate Breccia.—In many places, the porphyry and adjacent country rock are brecciated along zones of faulting and there is considerable intermingling of fragments. Where such breccia zones occur along the contact between conglomerate and porphyry, porphyry fragments intermingle with conglomerate boulders giving rise to porphyry-conglomerate breccia. A typical exposure of such a zone will grade from good conglomerate through a mixed zone of conglomerate and porphyry fragments into porphyry breccia and from there into massive porphyry.

Several specimens were collected from a porphyry-conglomerate breccia zone between conglomerate and quartz-feldspar porphyry 600 feet west of the Duparquet-Destor township line and 600 feet south of the Beattie road.

The fragments are generally less than one inch across and the brecciated

structure is visible only in thin section. In the field, the zone appears as an intimate mixture of porphyritic material with conglomerate pebbles and boulders. The phenocrysts are quartz and occur in the pebbles as well as in the matrix. No feldspar phenocrysts were observed in this zone.

The fragments, as observed in thin section, consist mainly of quartz-feldspar porphyry with a uniform fine-grained matrix, some conglomerate and some fractured and granulated quartz phenocrysts. The matrix between the fragments consists of quartz with a little sericite and is coarser grained than the groundmass of the quartz-feldspar porphyry fragments (Fig. 5). The larger crystals approach the size of phenocrysts and many are lath-shaped. These differ from typical phenocrysts found in the porphyry in that they have delicate lace-like borders and marginal zones containing many minute inclusions (Fig. 6). The quartz in these marginal zones is in optical continuity with the rest of the crystal. The lath-shaped quartz crystals are generally larger than the surrounding fragments. They contain a few inclusions of porphyry fragments and, in places, they penetrate and pass through breccia fragments, and hence are regarded as later than the fragmented portion of the breccia (Fig. 7). These quartz crystals are megascopically visible in both matrix and fragments of the breccia zones. The smaller quartz crystals of the matrix have indefinite wandering outlines and are also characterized by delicate lace-like borders, frequently crowded by minute inclusions. The matrix fills the fractures in the fragments and cements the granulated quartz crystals (Figs. 5, 6). There is no indication of crushing or granulation of the matrix in any of the sections of breccia examined.

Breccia zones exhibiting features similar to those of the zone just described may be seen in the north part of range IV, 2,400 feet west of the Destor-Duparquet township line.

Several other porphyry-conglomerate breccia zones on examination in thin section were found to exhibit some additional features. Such a zone outcrops on the west side of the Macamic road in lot 42. The north side of the outcrop is good conglomerate grading towards the south into a mixed zone of porphyry fragments and conglomerate boulders and then into massive quartz-feldspar porphyry.

In the thin sections examined, the fragments present are feldspar porphyry, quartz-feldspar porphyry, chert and crushed quartz granules. The matrix is similar to that in the breccias previously described. However, they are distinguished from the other breccia zones by the presence of numerous quartz veinlets, and veinlets containing carbonate with a brownish stain, hematite, pyrite and quartz. These veinlets clearly cut both the matrix and the fragments. Locally, the carbonate, hematite, pyrite, and quartz cements the breccia fragments.

Another example showing a different variety of this type of mineralization occurs in lot 37, 400 feet north of the railway in Destor township. At this locality an outcrop of conglomerate is in contact with quartz-feldspar porphyry to the north. The contact between conglomerate and massive porphyry is gradational over a width of 80 feet. Near the edge of the massive porphyry are angular inclusions of greenstone. These increase in number as the con-

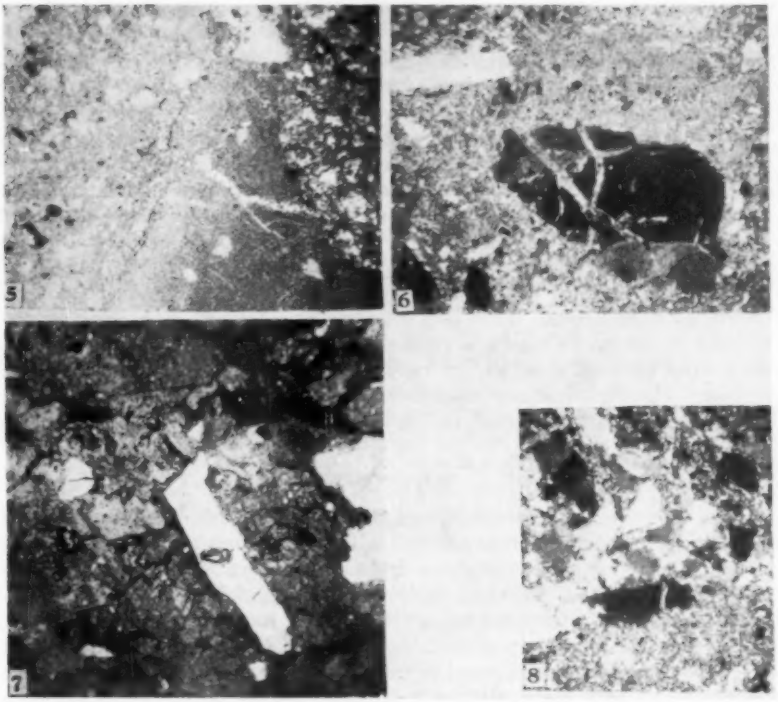


FIG. 5. Fragment of fine grained quartz-feldspar porphyry in silicified matrix of porphyry-conglomerate breccia. Quartz similar to that in the matrix seams the fragment. X-nicols, × 11.7.

FIG. 6. Matrix of porphyry conglomerate breccia filling fractures in quartz "phenocrysts" in the center of the picture. Note irregular shape and size of quartz grains and to upper left a quartz lath with lace-like inclusion rim suggesting growth by replacement. X-nicols, × 11.2.

FIG. 7. Quartz crystal in porphyry conglomerate breccia. Note upper part of crystal penetrating a fragment. Surrounding fragments are smaller than quartz crystal. X-nicols, × 22.8.

FIG. 8. Photomicrograph of crushed quartz phenocryst cemented by matrix of porphyry conglomerate breccia. X-nicols, × 56.

glomerate is approached. The porphyry assumes a brecciated appearance, and chert fragments, characteristic of the conglomerate, appear. Progressively nearer the conglomerate, the number of conglomerate fragments increase, jasper fragments also characteristic of the conglomerate appear and the amount of porphyritic material decreases. Finally, 80 feet away from the massive porphyry, the exposure is good conglomerate with a chloritic matrix. The boulders are composed mainly of angular fragments of chloritic

basalt with some chert and jasper pebbles. The matrix is chloritic and contains numerous quartz eyes.

In thin section the quartz-feldspar porphyry is well fractured. The fractures are filled with sericite, chlorite and quartz, and they cut the quartz and feldspar phenocrysts as well as the groundmass.

A specimen of the contact zone taken 50 feet from the porphyry was examined in thin section. The matrix surrounding the chert and greenstone fragments appears to be normal quartz-feldspar porphyry and is fractured to the same extent as the porphyry previously described. Well developed phenocrysts of quartz and feldspar are present. The quartz phenocrysts have deep, smoothly rounded embayments and stout crystal outlines. Over 90 percent of the feldspar content and less than 10 percent of the quartz content occurs as phenocrysts. This is a characteristic feature of normal quartz-feldspar porphyry. Chlorite is more abundant than usual, occurring as laths, the central portions of which are replaced by carbonate containing shreds of sericite. Minute veinlets of chlorite and carbonate fill the fractures, but the hematite and quartz observed in the breccia previously described is absent.

REPLACEMENT

The quartz-feldspar porphyry-sediment contacts have been shown to be gradational. The porphyritic material in the contact zones, although identical to the groundmass of normal quartz-feldspar porphyry and containing quartz phenocrysts, differs from the porphyry in that it contains no feldspar phenocrysts. This is a significant feature when it is recalled that approximately 40 percent of the normal porphyry is composed of feldspar phenocrysts whereas only about 1 percent is composed of quartz phenocrysts. If the contact zones represent part of an ancient regolith derived from the porphyry, as has been suggested by some previous workers, where are the feldspar phenocrysts? Why are they not present in the porphyry lenses in the contact zones as are the quartz phenocrysts? Furthermore the feldspathic material of the sediments in the contact zone progressively becomes more altered to white mica as the porphyry is approached. This micaceous alteration product and a few grains of detrital quartz occur as inclusions in the porphyry at the edges of the contact zones and in the porphyry material in the contact zones.

The inclusions of sedimentary material in the porphyry and the increasing alteration of the sediments as the porphyry is approached further preclude the possibility of the contact zone being of detrital origin. Rather, they are indicative of intrusion of the sediments by the porphyry and of reaction of a magmatic phase of the porphyry with the sediments.

Alteration of feldspar to white mica in a hydrothermal medium has been produced by Gruner (8) in the laboratory; consequently, some of the white mica found in the quartz feldspar porphyry was probably derived from the feldspathic material in the sediments. The quartz of the sediments occurs in places as inclusions in both the porphyritic material in the contact zone and the porphyry at the edge of the contact zone. However, the greater proportion has apparently been recrystallized and incorporated into the siliceous

porphyry groundmass. The chlorite in the sediments has also to some extent been incorporated into the porphyry. This is indicated by the fact that it is noticeably richer in chlorite near chlorite-rich faces of the sediments. The process of replacement that accompanied the introduction of the porphyry is apparently one of making over and finally of incorporating sedimentary material into the porphyry proper.

MODUS OPERANDI

After the crystallization of most of the feldspar as phenocrysts, the proportions of minerals that formed from the still liquid phase were enriched in quartz and white mica and the remaining accessory minerals, and impoverished in feldspar. The high proportion of mica observed in the thin sections indicates a "wet" phase of the magma at this stage. The original magmatic pressure consequently would have increased considerably and the magma would also be more fluid. The motivating force provided by the increased internal pressure is believed to have caused this mobile magmatic fraction to move out from the porphyry proper, forming an aureol of premeation in the country rock. This fraction was apparently too low in feldspar content to permit the formation of feldspar phenocrysts. The process of replacement preferred the matrix of the conglomerate and many localities may be seen where boulders of conglomerate are surrounded by a matrix partially or completely replaced by porphyry.

Subsequent to this period of replacement there was a period of brecciation that formed the porphyry conglomerate breccias. The phenocrysts in the partially replaced zones were crushed, as shown in Figures 6 and 8, and the porphyry and sediments were brecciated. However, there is no trace of brecciated structure in the matrix. On the contrary, the quartz, which forms over 90 percent of the matrix, has wandering, stringy outlines and fills the fractures in the broken quartz phenocrysts (Figs. 6 and 8). The larger quartz grains approach the size of the phenocrysts in the porphyry and are larger than the majority of the surrounding fragments. They contain inclusions of breccia fragments and penetrate and pass through fragments. Their lace-like borders, which are crowded with minute inclusions, indicate outward growth by replacement (Fig. 6). The phenocrysts in the porphyry proper, which contain the lace-like borders or growth rims, are believed to represent enlargements of normal phenocrysts by later quartz. Some of the larger quartz crystals in the breccia zones also undoubtedly represent enlargement of quartz grains, perhaps original fragments, to form "phenocrysts." It is thus apparent that a later period of replacement occurred after brecciation.

The replacing material differs from that of the primary porphyry in that it is composed almost entirely of quartz and is coarser grained (Fig. 5). The contrast between the grain size of two types of matrix is well illustrated in the quartzose material surrounding the large fragments in Figure 6.

This replacement is confined to, or near the margins of, bodies of quartz-feldspar porphyry. It is thus believed to represent a further continuation of the process of replacement by solutions emanating from the porphyry.

In addition to the high magmatic pressures prevailing, the porphyry was still subjected to the final stage of stress that implemented the preceding period of brecciation. This stress would provide an additional motivating mechanism whereby the residual solutions were squeezed out of the porphyry by a process of filter pressing, into adjacent breccia zones.

Following this period of replacement, veinlets composed of quartz, chlorite, carbonate, hematite and pyrite were formed. Locally these further replace the matrix around the fragments and are clearly later than the later siliceous replacement. The gold mineralization of the Duquesne, Golconda and Pitt deposits occurs in and adjacent to these veinlets. The close spatial relationship of these deposits to the porphyry and the presence of quartz, chlorite, carbonate, hematite, pyrite and gold in areas of siliceous replacement in and adjacent to the porphyry suggest a genetic relationship representing a further differentiation of the solutions derived from the quartz-feldspar porphyry.

THE THEORY OF PORPHYRITIZATION

The "phenocrysts" of quartz, which have been developed during the process of replacement, form the halos of quartz crystals that surround bodies of quartz-feldspar porphyry. The replacement accompanying their formation has been attributed to progressively changing solutions that are believed to represent stages in differentiation of a cooling porphyry magma. The modus operandi has been divided into five stages. Actual porphyritization is, however, confined to stages two, four, and to a minor extent the fifth stage.

First Stage.—During this stage the intrusion of the porphyry proper took place. The mode of intrusion was forceful. This is indicated by the presence of numerous inclusions of the country rock near the contacts, the numerous dikes found in the Keewatin complex, the occasional sill-like body of porphyry found within the Duparquet sediments and the bowing of schistosity where the porphyry intrudes into shear zones. Apparently crystallization had not begun in the porphyry at the time of intrusion as there is no flow structure in the phenocrysts along the contacts.

Second Stage.—The second stage began with the crystallization of the quartz-feldspar porphyry. The feldspar phenocrysts were the first to form and continued to crystallize until all but approximately 6 percent of the feldspar had formed. At this stage, magmatic vapor pressure had increased considerably and the residual liquid was sufficiently tenuous to permeate the porous arkose and arkosic matrix of the conglomerate under the high pressure existing at the time. The feldspar crystals were left behind by a process of filter pressing and the liquid followed a sinuous course replacing the sediment and swelling into lens-like forms where the nature of the rock permitted. Quartz crystallized as phenocrysts both in the lenses and in the porphyry, and the residual feldspar crystallized as fine crystals in the matrix. Finally, white mica, residual feldspar, and some of the quartz crystallized from the remaining liquid. These occur as a fine-grained mosaic that is characteristic of the groundmass of the porphyry. The country rock in the zone of soaking must have been heated to nearly the same temperature as the porphyry during the

period of soaking and replacement. If it had been noticeably cooler, the porphyry lenses would have been cooled too quickly to permit the formation of phenocrysts.

An alternative to the idea of direct crystallization from the magma in the formation of white mica is the possibility that it resulted from an alteration of the feldspar of the groundmass. In this case, the alteration would have resulted from deuteric reaction as there is no evidence of dynamic metamorphism. The feldspar phenocrysts are only slightly altered to white mica and the quartz phenocrysts are not fractured. The smaller feldspar grains in the groundmass would expose a greater area to attack explaining the preference of white mica for the groundmass and the lack of alteration in the feldspar phenocrysts. The alteration of feldspar to white mica would release quartz and enrich the groundmass in this mineral. However, the manner of formation of the white mica does not affect the general process of differentiation and porphyritization.

Third Stage.—The quartz-feldspar porphyry became increasingly rigid as crystallization proceeded. Eventually, minor movement along pre-existing faults resulted in brecciation. Boulders in the conglomerate were fractured along zones adjacent to the porphyry. Fracturing and brecciation provided channelways for solutions to penetrate the more impervious rocks.

Fourth Stage.—Subsequent to brecciation the residual liquid in the porphyry was predominantly silica with minor amounts of volatile constituents from which carbonate, some chlorite and sulfides and oxides of iron were finally deposited.

The fourth stage is marked by the introduction of these solutions into the breccia zones in the quartz-feldspar porphyry and adjacent rocks. Their introduction was aided by the remaining increment of stress that caused the preceding period of brecciation and which squeezed out the solutions by a process of filter-pressing. During this stage quartz crystallized but the other constituents remained in the fluid state. In the porphyry this resulted in the healing of fractures and in the healing of the brecciated phenocrysts by veinlets of quartz. These veinlets are locally terminated by, and have satellitic aggregates of irregular quartz "phenocrysts" crowded by inclusions with lace-like borders characteristic of growth by replacement. They are distinct from the primary quartz phenocrysts, which are clear and have knife-edge borders. Also present are growths of lace-like quartz surrounding large clear quartz phenocrysts. These commonly have a slightly different optical orientation from the clear nuclear portion and are attributed to the growth of quartz during this stage. In the brecciated rocks adjacent to the porphyry the breccia matrix is replaced by quartz and quartz "phenocrysts" occur in the replaced matrix and in fractures in the fragments. These have the characteristics of growth by replacement previously described.

Quartz "phenocrysts" were developed in this manner in the matrix of breccia zones and in fractures in the fragments and conglomerate boulders. This development of quartz "phenocrysts" by replacement occurred through a process that was essentially hydrothermal rather than magmatic.

Fifth Stage.—The fifth stage completes the crystallization of the porphyry and is marked by deposition in veinlets of chlorite, carbonate, hematite, pyrite and a small amount of quartz. Deposition of gold also occurred at this time. These effects are not as marked as those of stages two or four. There is some overlapping of this stage with the fourth stage as small amounts of carbonate, hematite and pyrite may be present in quartz veinlets of the fourth stage while quartz may be present in veinlets of the fifth stage.

SUMMARY OF CONCLUSIONS

Porphyritization is the development of a porphyritic and porphyroblastic texture in a previously formed rock by partial magmatic replacement and by metasomatism.

The "porphyry conglomerates" are the result of brecciation, replacement, and porphyritization, which took place during the solidification of the quartz-feldspar porphyry intrusives.

Porphyritization has been developed to varying degrees in other formations that occur in contact with quartz-feldspar porphyry. These are gabbro, chert, basalt, trachyte, agglomerate and older feldspar porphyry.

Porphyritization in arkose, graywacke, older feldspar porphyry and some phases of the quartz-feldspar porphyry is of the second stage. That in basalt was of the fourth stage along with subordinate second stage porphyritization. Porphyritization in boulders in the conglomerate, in gabbro and in chert, was of the fourth stage.

In the agglomerate, the "phenocrysts" are composite and are correlated with the fifth stage.

TORONTO, ONTARIO,
Mar. 25, 1958

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SCIENTIFIC COMMUNICATIONS

ALBITE PORPHYRIES AS A GUIDE TO GOLD ORE

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The goldfields of Ontario and Quebec in Canada, the Central Goldfields of Western Australia, the Kolar Goldfield of India and many others listed by Gallagher (3) are associated with albite porphyry dikes, which in different countries have been called keratophyres, albitites and alaskite porphyries.

Types of ore deposits genetically related to albite porphyries may be:

1. Auriferous silicified albite porphyry.
2. Auriferous quartz veins in albite porphyry.
3. Auriferous quartz veins in the country rock.
4. Auriferous sulfide lodes in the country rock.

The four different types of gold deposits represent stages in the differentiation of an albite porphyry magma, which in itself may be the end-stage differentiate of a basic magma.

This general occurrence throughout the world, with the principal deposits in Precambrian rocks, leads to the question "Is it possible to assess in a broad way an auriferous district on the evidence of one or two gold mines in which albite porphyries are found?". The writer is not thinking of the sparsely vegetated areas but more of the undeveloped tropical parts of the World where outcrop is largely obscured by the dense vegetation cover. Geological maps, in fact maps of any short, may be sketchy or completely lacking and consequently it is not easy to form an idea of the broad regional structures which might have a bearing on ore deposition.

If the albite porphyries are differentiates of a tholeiitic magma as postulated by Prider (7, 8) for some albite porphyries from Kalgoorlie, Western Australia, then the distance from a granite batholith will have no practical importance. Geological structures are of morre immediate practical significance. Throughout the Central Goldfields of Western Australia crossfolding has exerted a great influence on the localization of gold ore deposits. Geologists of the Geological Survey of Western Australia, notably H. A. Ellis, R. A. Hobson, and R. S. Matheson have had considerable success in recommending areas for prospecting by projecting crossfold axes to their intersection with "favorable beds."

Newhouse (6) has said "It is probable that if a syncline is mentioned as the major structural fold in a district, or if the ore deposits are described as

areally associated with synclines, that this areal association is closer to the synclinal axis than to the axis of any adjoining anticline." Dougherty (2) has shown that Kolar, Porcupine and Kirkland Lake are all related to synclinal structures. The Kalgoorlie syncline is an infold in the regional anticlinal structure. A scarcity of outcrops in the Coolgardie area twenty five miles to the west of Kalgoorlie has prevented the delineation of the detailed geological structure. The ore deposits are on the eastern side of a domal structure the core of which is occupied by granite. Nevertheless, from the sections available (5) the ore deposits seem to be localized in crossfolded synclines. The Santo Domingo gold mine on the eastern side of the Andes in Southern Peru is on the east limb of a northwesterly-plunging syncline (personal observation). The auriferous quartz vein which has been mined is associated with albite-bearing dikes.

Albite porphyry gold deposits are characterized by a simple mineral association. The commonest association of gold is with pyrite, pyrrhotite and arsenopyrite. Chalcopyrite, sphalerite and galena occur in small amounts and scheelite and gold tellurides are in some deposits.

The most striking similarity of the gold ore deposits related to albite porphyries is the high ratio of gold to silver in the veins and lodes. In Western Australian deposits the ratio of gold to silver is generally more than 9:1. This is particularly true of the Coolgardie Goldfield, which has produced practically no silver. Dougherty (2) reports that about 80 to 90 percent more gold than silver is present in the lodes of Porcupine, Kirkland Lake and Kolar. Exceptions occur, as for example at the Hasaga mine, Canada, (4) where the ratio of gold to silver is as low as 2.5:1. Derry (1) records that at the Canadian Malarctite mine "the gold-silver ratio in the porphyry being 6.28:1 whereas that in the graywacke is 1.9:1."

A simple mineral association and a high gold-silver ratio point to a genetic association and a high gold-silver ratio point to a genetic association with albite porphyries. Once this relationship is established the possibility of the four different types of gold ore deposits stated above can be suspected. In alluvial gold areas, exposures of albite porphyries similarly advise the search for more than one type of gold ore deposit. Regional structures suggest not only the minor controls in a district but may even suggest the presence of a large deposit.

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MOUNT PLEASANT,
WESTERN AUSTRALIA,
June 2, 1958

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DISCUSSIONS

ON THE OCCURRENCE OF URANIUM IN ANCIENT CONGLOMERATES

Sir: In this paper, ECON. GEOL., Vol. 52, No. 6, 1957, C. F. Davidson has again expressed himself strongly in favor of a hydrothermal origin for the mineralization of the Witwatersrand beds in general, and of the economic reefs in particular. Davidson's arguments are entirely on the mineragraphic side of the problem. In this connection he states: (p. 682) "The fact that the Witwatersrand shales are among the least radio active of the System is one of the most convincing contradictions of the placer theory." And on page 689, referring to Simpson's papers on radio logs of boreholes through the Witwatersrand System, he states: "They (the radio logs) do show clearly that the arenaceous beds are not less and generally more radio active than the argillaceous ones."

These statements do not agree with the results of numerous radio logs, carried out by the Mines Geological Department of Union Corporation Ltd., in the Kinross Area, Far East Rand and Klerksdorp Area. These show that Kimberley Shales are as radio active as large parts of the Kimberley Elsburg Quartzites and more radio active than Footwall Quartzites of the Kimberley Reef, and than Main Bird Quartzites between the Kimberley Shales and the Upper Bird Amygdaloidal Lava. Jeppe Shales are more radio active than Main Bird Quartzites. The Upper Shale Marker (or Vaal Reef Marker) is more radio active than large parts of the Gold Estate Quartzites and as radio active as the Vaal Reef Quartzites.[1] This agrees with Simpson's results in this area,[2] as shown on Plate V1. The sections of H. B. 6, H. B. 16 and H. B. 8 show the M. B. 1 section (= Kimberly Shale or Upper Shale Marker) as being more radio active than the Gold Estate Quartzites). Lower Witwatersrand sections show that in general the quartzites and shales are equally radio active, with variations both ways.

It would be more correct to say that the Main Bird Quartzite is among the least radio active formations of the System.

Davidson's statement (p. 673): "In any single sedimentary cycle within the System, the radio activity progressively decreases as one passes from conglomerates through quartzites and argillaceous quartzites to shales." finds also no confirmation in our radio logs.

Davidson's objections to the suggested glacial origin of the Witwatersrand conglomerates are not convincing. In order to be able to throw out this

[1] These results have not been published. This statement is given with the permission of the management of Union Corporation Limited.

[2] D. J. Simpson, Ph.D., Correlation by means of radio activity logging in the Witwatersrand System in the Klerksdorp Area, Trans. Geo. Soc. S. Africa, 1952, p. 33.

suggested glaciation, he acts rather drastically and casts serious doubt on the accepted tillites in the Lower Witwatersrand System (and of the Transvaal System tillite also, I suppose) because of their oligomictic character. If an ice sheet has not moved over a large variety of rocks, one cannot expect that it will produce a polymictic moraine. It is not necessary to postulate a time interval of upwards of 20,000,000 years for the deposition of 35,000 feet of Witwatersrand sediments. The thickest succession, that of the Central Rand, is only 25,000 feet thick and a large part of this succession consists of conglomerates and coarse quartzites. It seems improbable that this succession would have been deposited at an average rate of one inch of sediment every 50 years (for 35,000 feet), or every 65 years (for 25,000 feet). A small fraction of this time should be closer to the truth.

All the difficulties Davidson experiences (p. 677–678) in explaining uniform placer mineralization of thin conglomerates by alluvial processes fall away if he replaces alluvial processes by diluvial ones. The uniform distribution of material from point sources many miles away, the physicochemical conditions which he can neither envisage nor explain in connection with the transportation of ore minerals—all this can be explained if it is assumed that inland ice sheets transported the ore minerals from their source to their present location.

Davidson does not describe in detail how he thinks that the epigenetic mineralization of the Witwatersrand beds took place. So far no Rand geologist has been able to do so either and numerous are the arguments brought forward during the past decennia against the epigenetic processes whereby mineralizing solutions should have followed in a miraculous way few horizons, which from a point of permeability and of persistence over large areas would be indistinguishable from other horizons. These other horizons were left out by these solutions and remained practically barren. If permeability of a conglomerate played a role at all in the mineralization process then the degree of permeability should be a point of consideration for a mining geologist when determining the reef. This is not the case. After the reef has been lost on a fault or dike, the reef, when picked up again, is determined by its place in the succession. Points of consideration are: components, color, degree of cleanliness, size of particles, silicification, etc., but never permeability. The reef may be a thin grit or may be entirely absent, the hanging wall lying directly on the footwall. Well developed leader reefs may occur few feet away in hanging wall or footwall, offering far better travelling ways for solutions than the reef itself, but they will be barren or very poorly mineralized and any worth while mineralization will be at the reef horizon and this is entirely determined by its place in the succession and by nothing else. This certainly points to a process of mineralization that took place at the time the reef was deposited. That a mineralizing solution was able to follow such a horizon over dozens of miles would be nothing less than a miracle.

Moreover, if Davidson wants to put the time of mineralization into the period of post Transvaal earth movements (p. 691) then he is confronted with the difficulty that the Witwatersrand formations were already solidified

to such a degree by that time, that the conglomerates could not have offered travelling ways at all. Many of the pre-Ventersdorp and Ventersdorp age faults could have done so, but none of these show any gold-uranium mineralization. The conglomerates and sandstones of the Transvaal System would have been the most permeable horizons available, but they are barren, except for the sporadic mineralization of the Black Reef at the base of the system.

Davidson has paid no attention to the geological conditions under which the selective epigenetic mineralization had to take place. Whatever mineragraphic arguments can be brought forward against a sedimentary origin, these arguments seem to be insignificant compared with the geological arguments against an epigenetic origin.

The "Deus ex machina" is clearly on the other foot.

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March 20, 1598

GEOCHEMISTRY, CRYSTAL STRUCTURE AND MINERALOGY OF THE SULFIDES

Sir: In vol. 52, p. 755–774 of this journal an interesting paper on the geochemistry, crystal structure and mineralogy of the sulfides was presented by Virginia Ross. I feel that the interpretation given in this paper of an earlier paper by Frueh (2) on the covalent-metallic nature of bonding in sulfides is very misleading, and that the interpretation given the data of some other authors on the electrical resistivity and reflectivity of sulfides is definitely erroneous.

In her paper, on page 757, Dr. Ross states that Frueh's attempt to utilize Brillouin zone theory rests on the assumption that the bonding electrons are essentially free. From the discussion following this point and from the discussion of "Luster and Reflectivity" on page 770, it is evident that by "free electrons" Dr. Ross refers to the conducting electrons. In the paper by Frueh (2) the term "free electron" was used, perhaps unconventionally, to distinguish between the tightly bound electrons that are confined to states about the nucleus, and the outer shell electrons that are free to be used for bonding. This followed from the simplified development of the theory given by Hume-Rothery (4) which started by considering free electrons in a box before showing the restrictions placed upon these electrons by the periodic distribution of potential in a crystal. In the case of the alkali metals the valence electrons that comprise the Fermi body do not fill the first Brillouin zone and consequently they are also conducting electrons.

It was pointed out by Frueh that a great many of the sulfides could be considered as semiconductors. This means that they would behave as electrical insulators when at low temperatures and of perfect purity. A "Brillouin

zone," or band of allowed electronic energy levels known as the valence band, would be completely filled. An energy gap of forbidden energy levels would exist between this band and the conduction band. There would be essentially no electrons in the conduction band—no conduction electrons, or those referred to conventionally as "free electrons"—and yet the Brillouin zone theory gives us a reasonable picture of the restrictions placed on the energy states of the valence electrons.

These sulfides become conductors for one or both of the two following reasons:

1) They become intrinsic semiconductors if the energy gaps are small enough that with increasing temperatures the electrons at the high energy edge of the valence band will be given enough energy to jump the forbidden energy gap and enter the conduction band. These electrons will become conducting electrons and additional "hole" conductivity will take place due to the vacancies in the electronic states of the valence band.

2) They become extrinsic semiconductors by the addition of impurities that either accept electrons (*p*-type semiconductors) from the valence band, thus allowing "hole" conductivity, or donate electrons (*n*-type semiconductors) to the conduction band.

The energy gap between the zones, as well as the energy levels of the impurity atoms, can be determined experimentally from resistivity vs. temperature measurements. The utilization of the band structure to explain certain properties of the sulfides does *not* rest on the assumption that bonding electrons are conducting electrons.

On page 771 Dr. Ross utilizes a graphical plot of resistance vs. reflectivity to show a rough correlation between luster and "free electron" character. For resistance values (ordinate), data were obtained from Harvey's (3) measurements made on the surface of polished sections. Harvey's intention was to investigate the possibility of utilizing resistivity as an aid to identification, and there was no uniformity as to size of sample, shape or orientation, purity, homogeneity, presence of interfaces or other factors that would effect the resistance. Dr. Ross states she used Harvey's mean values, and in many cases she does just this. In some instances, as in galena, chalcopyrite, and pyrite, the minimum value instead of the mean value is plotted. This makes a considerable difference in the location of the point on the graph. For example, Harvey lists 17 samples of galena giving readings varying from 10^{-2} to 10^{2} ohms, and 7 samples with values in the vicinity of 10^{3} ohms. This should yield a mean value somewhere in the vicinity of 0 to 10 ohms, but the point for galena appears on the graph at 10^{-2} ohms. Equally extreme errors are made for chalcopyrite and pyrite.

Even if these errors had not been made it is doubtful that the mean value would in any way relate to the "free electron" character of the bonds. In most of these materials the conductivity at room temperature is extrinsic due to the occupation of impurity levels as explained above. Certainly the wide variation of resistance values seems to indicate this. In the case of galena (PbS) and pyrite (FeS_2) the recent works of Bloem, Kröger, and Vink (1),

Scanlon (8), Marinace (5), and Sasaki (7), clearly show that the intrinsic range, when the bonding electrons are involved in the conductivity, is not reached in these minerals until over 200° C. Therefore to interpret the room temperature conductivity in terms of the bonding electrons rather than in terms of impurity levels has little meaning.

Plotted against resistance is the reflectivity (absissa) obtained from Short's (9) measurements made with white light (spectrum not noted). If we had some notion of the variation of reflectivity with wave length of light we might obtain some information about the surface states, and from optical absorption data further information about the electronic states within the mineral could be obtained. I fail to see, however, how Short's data, again obtained for the purposes of identification, can be used for the interpretation it has been given.

I am afraid I disagree with Dr. Ross on several other points. For example, in Table 3, "The structural classification of sulfide minerals," argentite and α-naummannite head the list of cubic close-packed structures of the anti-fluorite type. Until there is some evidence presented to indicate that Rahlfs' (9) determination of the body-centered cubic arrangement of the sulfur atoms in these minerals was incorrect, I feel they should not be classified under anti-fluorite structures. I also would question the placement of α-stromeyerite and α-eucairite under anti-fluorite type derivatives.

Dr. Ross's paper serves to point out the tremendous amount of data within the scope of present-day science that still must be obtained before a modern interpretation can be given to the geochemistry, crystal structure and mineralogy of the sulfides.

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GEOCHEMISTRY, CRYSTAL STRUCTURE AND MINERALOGY OF THE SULFIDES

Sir: In the application of Brillouin zone theory to the system: Cu—Fe—S (1), Dr. Frueh states (p. 84): "If we assume that all the outer shell electrons of the elements of this ternary system could contribute to the free electrons composing the Fermi body, then copper would yield one free electron, iron would yield three and sulphur six." Apart from the fact that the valence states of the metals are to be questioned and the bonding electrons are not completely free, the simple band model does not provide an accurate description of the electrical properties of compounds possessing a high degree of heteropolar and homopolar character. The mere knowledge of the form of the Brillouin zones and the number of so-called "loosely bound" electrons required to fill them in the cases cited by Dr. Frueh tells us *at most* that the number of "effective" free electrons, $N_{eff.}$ is very much smaller than the actual number of valence electrons.

The exact forms of the energy surfaces and magnitudes of the discontinuities across the various boundaries depend upon the fields within the minerals. The width of the forbidden zone, $\triangle E$, increases in proportion to the difference between the maximum and minimum electronic potentials in the periodic lattice. Consequently the electronic carrier mobility in such compounds decreases with increasing ionicity of the bonds. Since the band structures of zincblende-type crystals in isoelectronic sequences are still incompletely understood, it may be unwise to speculate on chalcopyrite ($CuFeS_2$) which is not only in diagonal sequence, but polymetallic with at least four different valence states in total. A recent survey of the theories regarding band structure and the ionic component in Group II-Group IV compounds is to be found in Solid State Physics by Welker and Weiss (2).

The equations for X-ray diffraction and Brillouin zone boundaries are very similar. In monatomic solids the electronic potential is approximated by the sum of the contributions of individual atoms, so that energy discontinuities are small for wave numbers for which the x-ray structure factors are small and large when the structure factors are also large. *In polyatomic solids, such as the sulfides, Dr. Frueh examines, the assumption is made that the scattering powers for low-energy electrons and for X-rays are proportional for the different atoms.* This is not necessarily true.

While not so elegant in style, a similar argument can be advanced that the face-centered-cubic-structure-type in the minerals chalcopyrite, bornite, and their solid-solutions at elevated temperatures is maintained by a constant ratio of valence electrons to cell volume (atoms plus holes). The deviations from ideal stoichiometry and lattice defects found in bornite, digenite and stromeyerite, also considered by Dr. Frueh (3), are probably special cases involving the substitution of one Cu^{+2} ion plus one hole for two Cu^{+1} ions. The electronic energy scheme then contains levels due to hole imperfections and levels of the substituting atoms. In such cases both levels are close to the conduction band.

Resistivity and Reflectivity of the Sulfides.—Dr. Frueh interprets room-temperature conductivity in the sulfides as being entirely due to impurity levels, but in reference to stromeyerite (3), he has stated to the contrary that: "the bonding is undoubtedly somewhat metallic in nature" (p. 304). Appreciable metallic character (as denoted by luster, opacity, electrical- and thermo-conductivity, long deformed bonds, etc.) is manifest in those pure sulfides, arsenides, antimonides and tellurides which are composed of atoms which differ only slightly in electronegativity. The resulting bonds are of mixed character; the presence of "subsaturated" valence states gives rise to some incompletely restricted electrons. In many sulfides, the metallic properties are enhanced by or associated with the presence of impurities or the stoichiometric excess of atoms. Since the pure mineral is a rarity, the influences of impurities cannot be ignored. *However, there are apt to be limits on the type and concentration of impurity and the resultant energy gaps associated with a specific structure and electronic environment.*

In "Geochemistry, crystal structure and mineralogy of the sulfides" (4) the author has obtained a "rough correlation" between resistivity ("free-electron" character) and reflectivity (p. 770–771). Since the plotted values of the resistivity and reflectivity of the individual sulfides were obtained from totally different specimens, studied by different investigators, it is small wonder that any correlation could be found. Certainly, more elaborate measurements on concentration levels of impurities and defects are necessary to establish the causality of the relationship. Mean values of the resistivities and reflectivities were used to obtain this curve. It can be extrapolated to include the majority of points which lie in the region of resistivity greater than 0.1 ohm and reflectivity on the order of twenty percent. Owing to tremendous differences of three to four orders of magnitude, the limiting values were plotted for galena, pyrite and chalcopyrite. Measurements of the resistivities of these same three minerals by Kerr and Cabeen (5) are in close relative agreement with the minima shown in the graph. Furthermore, the same average value has been obtained by Sasaki (6) from very accurate measurements on natural pyrite crystals from Japan. The perfect natural pyrite crystal without impurity energy levels does not exist, so the high (room-temperature) intrinsic resistivity is of no practical importance to the mineralogist.

The precise interpretation of the absorption bands in the sulfides as a function of wavelength is a task for the solid state physicist, and will depend on accurate measurements in the region of short wavelength close to the ultraviolet. For the simple metals the critical wavelength between absorption and reflection can be correlated with $N_{eff.}$, the number of "effective" free electrons. In the sulfides, the sum of the photoelectric contributions of transitions from d, s and p electrons in their respective energy states are obtained in the visible region. Furthermore, most laboratory identifications involving reflectivity can be performed with the use of white light.

Structural Classification of the Sulfides.—The single structural classification of the sulfide and sulfosalt minerals (4) has been attempted with much data obtained since the publication of the last edition of Dana's System of Mineralogy, vol. 1 (7). The arrangement of the classification is sufficiently

basic to permit the inclusion of newly-determined structures as derivatives of the prototypes. The results of highly coordinated investigations of the sulfosalts (from the laboratories of E. Hellner (8)) are further proof of the close structural relationships between the chemically-simple and complex sulfides. The parameter $f_1 = \frac{Ag + Pb + Sb(Bi)}{S}$ for a specific sulfosalt, is indicative of the degree of similarity to the galena or stibnite structures. The ratio $f_2 = \frac{Ag + Pb}{Sb(Bi)}$ is directly related to the percent of octahedral or pyramidal bonding in the structure. Closely related to the galena structure are: freieslebenite, diaphorite, miargyrite, andorite, meneghinite, fizelyite, ramdohrite, cosalite, gratonite, jordanite, and geocronite. Derivatives of the stibnite structure are gladite, lindstromite and hammarite. Intermediate-type chain structures are berthierite and galeno-bismuthite. By choosing different crystallographic axes, such as (110) or (111), most of these structures can be more easily identified with the simple galena type. Similarly, the diagonal axes of the body-centered argentite and α-naumannite have been chosen for comparison to the face-centered anti-fluorite-type structure. The metal or anionic lattices of eucairite and stromeyerite are fundamentally related to these face-centered arrangements and the metals are in approximately tetrahedral coordination by sulfur (selenium). The high-temperature modifications of these minerals are included in the new classification as probable derivatives of the anti-fluorite-type on the supposition that the metal atoms are statistically distributed over the interstitial sites as they are in the simple di-metallic sulfides, selenides and tellurides investigated by Rahlfs (9).

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Economic Geology
Vol. 53, 1958, pp. 765–769.

REVIEWS

Zinn und Wolfram. Vol. II of: The metallic raw materials, their deposits and economic significance. By F. AHLFELD. Pp. 212; figs. 29; tables 33. Ferdinand Enke Verlag, Stuttgart, 1958. Price, DM 48.

Professor Friedenburg, the editor of the series "The Metallic Raw Materials," which was started many years ago by the late Paul Krusch, has found in Dr. Ahlfeld a highly qualified geologist with a life time experience in tin and wolfram geology and mining.

Accordingly, the treatment of the subject is thorough and up to date. The first half of the book deals with tin and the second with wolfram. Each of these sections is further subdivided into two parts, a general one (about 40 p. each) and a special one. In the general part, Ahlfeld discusses the properties of tin and wolfram, respectively, the history, the occurrence and genesis (geochemistry, minerals, and deposits), the recovery (mining, ore dressing, smelting, and refining), the use, markets, prices, reserves and future, and statistical data.

In the special part (about 70 p. each), the author offers a collection of excellent, brief, and concise reviews on the tin and wolfram districts according to countries. Each sub-chapter includes numerous references and it is easy to see that the author has always attempted to include the very latest information. The bibliography thus includes over 1000 references and many of our textbooks on Economic Geology can be brought up to date by comparison with Ahlfeld's treatise. A list of about 200 mines adds also to the value of the book.

Like the previous volumes of this series, the publisher has again made a successful effort to offer clear and neat figures and remarkably few printing errors are found.

G. C. AMSTUTZ

MISSOURI SCHOOL OF MINES
AND METALLURGY,
June 24, 1958

Vorratsberechnung Mineralischer Rohstoffe. (Calculation of Ore Reserves.) Translated from Russian to German by A. P. PROKOFJEW, edited by O. OELSNER, FREIBERG, and E. PEUKERT. Pp. 164; figs. 45; tables 20. VEB Verlag Technik, Berlin, 1956. Price, 10.-DM.

For a number of years the publisher VEB Verlag Technik in Berlin has edited a series of translations of outstanding Russian textbooks. Among these were Betechtin's Mineralogy, Beljankin's and Petrow's Optical Mineralogy, Saukow's Geochemistry, Sawarizki's Petrology, and Wachromeiew's Ore Microscopy. Perhaps the most important one of these translations is, for economic geology, the one of Prokofjew's textbook on the calculation of ore reserves.

The author introduces first the basic terms used for the calculation of ore reserves. In the second chapter he deals with the accuracies of such calculations and discusses in detail the possible errors and their influence on the results. He also introduces numerous correction coefficients. In the third chapter, which might

be called the "pièce de resistance" of the whole book, he then treats the methods of ore reserve calculations. He mentions twenty different methods used at different times and in different countries. He shows that most of these various methods, which at first glance look quite different, can be classified into very few basic types with their variations. All methods appear to be based on the basic elements of any calculation of reserves: the volume, the specific gravity, and the calculation of the average grade.

The author recommends only three basic methods, out of the large number of methods discussed in the third chapter. They are: 1) the profile method, 2) the black method (mining blocks), and 3) the method of "geological blocks," which he describes in the fourth chapter. The role of the geological type of an ore deposit for the calculation of ore reserves is emphasized and illustrated, with a number of actual examples.

The coverage of the English literature is practically lacking. None of the excellent papers which have appeared in the last fifteen years in Australia, in the United States and in England are mentioned. If one wants to read a brief and excellent comparison of Russian, European, and American methods one should refer to Vogel's article "Analysis of ore reserve classifications used in different countries and a few suggestions regarding such classifications with special reference to the mining of vein deposits" (Freiberger, Forschungshefte, C10, 1954, p. 5–32).

Prokofjew's book is nevertheless very interesting and can be highly recommended to anyone who likes to deepen his knowledge and anyone who believes that people in other countries might also have some good ideas. The clear figures and tables add much to the quality of the text which is written in clear easy German.

G. C. Amstutz

Missouri School of Mines
and Metallurgy,
June 24, 1958

Geologiya Sovetskoi Arktiki [Geology of the Soviet Arctic]. Edited by F. G. Markov and D. V. Nalivkin. Pp. 520, with colored geological map in four sheets, scale 1:2,500,000. Gosgeoltekhizdat, Moscow, 1957. Price 37 roubles. (Russian only.)

This noteworthy work, published to mark 40 years of Soviet studies in northern lands, is a collection of fifteen articles, from ten to 110 pages in length, each reviewing present knowledge of a sector of arctic Russia. The territories described include principally the archipelagos of Franz Josef Land, Novaya Zemlya, Severnaya Zemlya, the New Siberian Islands, and Wrangel Island, together with the vast mainland districts of the Polar Urals, Taimyr and other sectors of northern Siberia, the lower reaches of the river Lena, and the northern part of the Verkhoyansk mountains. In each case a brief statement on topography is followed by a lengthy account of the stratigraphy, volcanism and tectonics, with a brief note on mineral deposits. No references to earlier works are listed, there is no index, and the volume is devoid of illustrations. Among the mineral occurrences of importance are the platiniferous copper-nickel ores of Norilsk, the polymetallic mineralization of northern Verkhoyansk and the diamantiferous kimberlites of Yakutia. The geological map is excellently printed, with an index of 90 formations. On it is given the location of some 34 pipes of kimberlite.

C. F. Davidson

University of St. Andrews,
Scotland,
June 16, 1958

Minéraux d'uranium du Haut Katanga. J. F. VAES and C. GUILLEMIN, photography by A. DESTAS. Pp. 81; pl. 27. Les Amis du Musée Royal du Congo Belge, Tervuren, Belgium, 1958. Price, 240 Belgian francs (approx. $4.80).

This little book contains an outline by J. F. Vaes (pp. 7–20) of the mode of formation and associations of the secondary uranium minerals of Katanga, followed by succinct summaries, each about a page long, by C. Guillemin of the crystallographic, x-ray powder, optical, physical, and chemical properties of 33 secondary uranium minerals. These will be useful; the book is outstanding, however, for the 27 photographs in color by A. Destas, made at magnifications of × 7 to × 125. In brilliance of color, faithfulness of hue, and excellence of reproduction, these are the best photomicrographs of minerals I have ever seen. Anyone interested in minerals will get much pleasure, as well as profit, from the book.

MICHAEL FLEISCHER

The Law of Water Allocation in the Eastern United States. Edited by DAVID HABER and STEPHEN W. BERGEN. Pp. 643. Paper cover. Roland Press Co., New York, 1958. Price, $7.50.

This thick volume results from the papers and proceedings of a symposium held in Washington, D. C., October, 1956, sponsored by The Conservation Foundation, which is an independent American organization established to promote greater knowledge about the earth's resources. "This symposium; consisting of 12 contributing authorities and 52 participants and observers, was arranged to consider the problem of increasing conflicts of interest over water rights in the eastern United States, and the need for revising existing laws to meet changing conditions in the usage of water resources."

There are 9 groups of subjects, containing 13 papers, each group being followed by discussions. The water laws of Massachusetts, North Carolina, and Michigan are considered as case histories. There is a review of certain types of water use in Western law and practice, and some specific studies of certain areas. Other subjects covered are water allocation, economic and policy aspects of water use, technical aspects of water management, and legislative alternatives. The book includes contributions from the fields of law, economics, engineering, hydrology, administration, and viewpoints of major water-consuming interests.

The book is full of information for engineers, conservation commissions, legislatives, law teachers, public health departments, and geological surveys.

A Review of Commonwealth Raw Materials, 37th Report. Vol. I. Pp. 288. The Commonwealth Economic Committee, London, 1958. Price, 15s.

This well organized review takes up the production and consumption of raw materials in the Commonwealth countries, and the Commonwealth sources of energy. There are also chapters on Commonwealth trade, on the prices of raw materials, some trends and developments, and a Summary and Commentary chapter. Five appendices give notes on statistics, lists of raw materials and other considerations.

"The object of the enquiry is to show against the background of world conditions the post-war development . . . in the production, consumption, and trade of Commonwealth raw materials, and to compare the post-war with the pre-war condition." About two-fifths of all imports into the United Kingdom are of raw materials, and half of these come from Commonwealth countries. Thus the raw materials of the Commonwealth countries play an important role in the economy

of the United Kingdom and of the individual countries, in many of which the raw materials dominate in the net value of all production.

An interesting table on page 5 gives the indices of free world production of the important raw materials, using 1935–1937 as a base, for the years 1949 through 1955. Another table (p. 7) shows the increases and decreases of the Commonwealth share of free world production in 1955 as compared with 1949–54.

The various products considered are metals and ores, non-metallic minerals and organic raw materials. These are reviewed not only for the Commonwealth countries but for the free world countries also. Chapter 7, on Some Trends and Developments, gives interesting thoughts on the growth in use of certain materials, political effects, improved scientific methods, substitution, changing patterns of trade, and new sources of energy.

The book is a valuable contribution to knowledge of the sources of raw materials throughout the world and of the Commonwealth countries in particular, and will serve as a ready reference.

A Guide to the Known Minerals of Turkey. Revised by C. W. RYAN. Pp. 199. U. S. Operations Mission to Turkey, Ankara, Dec., 1957.

The original of this work consisted of some 2,200 entries; this revision contains about 3,300 descriptions, most of the increase being due to the inclusion of entries of copper, lead, zinc, and iron. This edition also includes, where available, descriptions of the geologic setting and the theories of origin of the deposits. Much of the material came from the library of the Maden ve Tetkik Arama Enstitusu in Ankara. Mineral maps compiled in connection with this study on a scale of 1:800,000 are available in the office of the mining advisor to E.C.A. in Ankara.

The deposits are divided into four groups: Group A, non-ferrous and minor metals; Group B, iron, ferro-alloys, metallurgical and refractory materials and mineral fuels; Group C, ceramic, industrial and manufacturing materials; Group D, chemical, fertilizer, abrasive and miscellaneous materials. Under each metal or mineral are given the localities where found, with a few lines for the unimportant ones and several pages for the chief deposits. Chromite has 16 pages. Past and present production and reserves are given.

The book is an excellent reference to all the past and present and potential mineral resources of Turkey.

The Prentice-Hall World Atlas. Edited by JOSEPH E. WILLIAMS. Maps 158, index pp. 26. Prentice-Hall, Inc., New York, 1958. Price, $9.00.

This handy atlas, size 8½ × 12 inches, is divided into three sections, (1) the Systematic Geography of the World, (2) World Economic Maps, and (3) Physical Maps of the Continents with separate maps of their geographic regions. The first section includes world maps showing political divisions, population, religions, climate, precipitation, soil groups, geology, tectonics, vegetation, and agricultural regions. The second section of world economic maps depicts the world distribution of 24 minerals, 34 agricultural products, sheep, cattle and swine. The third section includes most of the book, continent by continent. First are given continental geology, population, ethnographic, vegetation and land use, and economic. These are followed by maps of the individual regions showing elevations by color, principal geographic features, railroads, and cities indicated by population designations. An index of some 20,000 entries permits one to locate any place readily.

The new merged color, shaded relief maps are very striking and give an excellent three dimensional effect. The whole atlas is beautifully made and is one that every family and office should have.

Hydrogéologie, 2nd Edition. By P. FOURMARIER. Pp. 294; figs. 164. Masson and Cie. Paris, 1958.

This new edition, following its predecessor of 1939, has been brought up-to-date and considerably enlarged. It includes also an annex on the principals of modern procedures of purification and waters by Edmond Leclerc of Liege. The 18 chapters of the book are divided into three parts: (1) bodies of normal waters, (2) catchment of normal waters, and (3) thermal and mineral waters.

The first section takes up the cycle of waters, surface and subterranean waters, water tables, circulation, aquifers, and artesian waters. The second section deals with the qualities, contamination and purification, utilization, and hydrologic conditions in arid countries. The third section goes into bodies of thermal and mineral waters, their properties and utilization.

The various subjects are thoroughly treated and well illustrated. This new edition is the outstanding book in the French language on this subject.

Landscapes of Alaska—Their Geologic Evolution. Edited by HOWELL WILLIAMS. Pp. 148; plates 23; figs. 3; maps 5. University of California Press, Berkeley, 1958. Price, $5.00.

The timing of this book is almost perfect. A fortnight after its appearance Alaska was voted to be a state of the Union, and people are avid for information of this new state. This book will meet their desires, for written in language readily understood by the layman, it portrays this vast and scenic land, and dramatic photographs and maps give it realism.

The work is the product of eleven authors, mostly members of the U. S. Geological Survey. Bradford Washburn, Director of the Boston Museum of Science and well known Alaska explorer, supplied most of the photographs, and oil and mining companies helped finance it. Fifteen national geographic areas are described and pictured—the islands, fiords, rivers, mountains, glaciers, tundra and permafrost. Their geologic evolution is traced. The people are also considered—the Eskimo, the Indian, and the Sourdough. The mines, industries and communications also receive attention.

Alaska is a large country, much larger than Texas, extending from the frigid Arctic shores to the mild climate of Southeastern Alaska. It contains the highest mountains in North America and has more glaciers than any other area.

Those who wish to know more about Alaska or plan to travel there should become acquainted with this book.

BOOKS RECEIVED

JAMES M. ALLEN

Modern Russia—an Introduction. JOHN LONG. Pp. 180. Philosophical Library, New York, 1958. *A handy and interesting little book dealing with the physical setting, the peoples, the one-party system, the planned economy, Soviet Russia in the world today, and appendices.*

Statistical Summary of the Mineral Industry 1951–1956. Pp. 371. Overseas Geological Surveys, Mineral Resources Division, London, 1958. Price, £1, 7s, 6d. *Production, exports, and imports of all metals and minerals of all countries.*

Manual of Scientific Russian. THOMAS F. MAGNER. Pp. 102. Burgess Publishing Co., Minneapolis, Minn., 1958. Price, $4.60. *Serves as a textbook for reading scientific Russian and as a reference book for Russian scientific literature. Gives the structure of Russian, vocabulary, techniques of translation, and Russian texts and glossary.*

Oil and Gas in the Four Corners. PAUL J. KUHN, ed. Pp. 298. National Petroleum Bibliography, Amarillo, Texas, 1958. *A story of the development of 172 oil and gas fields in the Utah, Colorado, New Mexico and Arizona, Four Corners area, of some 70,000 square miles ranging in elevation from 3,500 to 13,500 feet in elevation. Structure, stratigraphy, development and details of each of the fields.*

Coal Exploration Records of Boreholes. Pp. 267; figs. 4; tbls. 7. Price, 3 sh. Geological Survey of Bechuanaland, Lobatsi, 1957. *Logs of 50 holes drilled into the Karoo of eastern Bechuanaland. Analyses of coal samples.*

Iniciacion de la Investigacion de Pastos y Forrajes en Colombia. LOY V. CROWDER. Pp. 7. Ministerio de Agricultura de Colombia, 1958. *The agricultural possibilities of Columbia.*

Studies of Devonian Algae. J. H. JOHNSON and KENJI KONISHI. Pp. 114; pls. 26; tbls. 7; maps 6. Price, $1.50. Quarterly of the Colorado School of Mines, Vol. 53, No. 2. Golden, 1958. *A compilation of known data about rock building algae of Devonian age.*

Report of the Director of Geological Survey 1956–57. Pp. 20; tbl. 1. Price, 2 sh, 6 d. Government Printing Dept., Publication Branch, Accra, 1958. *Summary of the activities of the survey during 1956–57. Brief notes on the origin of manganese deposits, petroleum exploration, geophysical and geochemical prospecting and diamond drilling.*

Well Location County Maps of Indiana Showing Total Depth of Wells—Polyconic Projection. Scale, 1 inch to the mile (approx.). Indiana Geological Survey, 1958. *Petroleum exploration maps of each county showing oil and gas wells, dry holes, abandoned holes, and wells in a single pool.*

Conodonts from the Glen Dean Formation (Chester) of the Illinois Basin. C. B. REXROAD. Pp. 27; pls. 6; figs. 14; tbl. 1. Illinois State Geological Survey, Rept. of Invest. 209, Urbana, 1958. *Detailed description of the conodont fauna of the Glen Dean formation. Seven new species recognized.*

Symposium Sobre Yacimientos de Manganeso, Tomo III. America del Sur. The Manganese Deposits of Minas Gerais, Brazil. J. VAN N. DORR, I. S. COELHO and A. HOREN. Pp. 67; figs. 6; tbls. 5. International Geological Congress, Mexico, 1956. *Geology, mineralogy, reserves, and location of deposits. Ore formed from low grade protore by processes of supergene enrichment. In English.*

Progress Report on Clays of Montana. U. M. SAHINEN, R. I. SMITH, and D. C. LAWSON. Pp. 40; pls. 2; tbls. 6. Montana Bureau of Mines and Geology Inform. Circ. 23, Butte, 1958. *Location, description of clays and deposits, summary of laboratory testing data.*

Estudos, Notas e Trabalhos. Bol. 24. Pp. 83; pls. 7; figs. 14; tbls. 6. Serviços de Indústria e Geologia, Lourenço Marques, 1957. *Eight papers—including descriptions of pegmatites, age determination, carbonatates banded ironstones.*

Annual Report—Nyasaland Protectorate, 1957. Pp. 40; figs. 3. Price, 5 sh. The Government Printer, Zomba, 1958. *Summary of activities for 1957 including areal geology, carbonatite and ring structure investigation and ground water studies.*

Cretaceous, Paleocene, and Lower Eocene Geologic History of the Northern Mississippi Embayment. R. G. STEARNS. Pp. 23; figs. 21; tbl. 1. State of Tennessee Dept. of Conservation, Div. of Geology Rept. of Investigations 6. Nashville, 1958. *Description of mixed marine and non-marine beds resulting from a single sedimentary cycle.*

On the Genesis and Mineralization of the Tungsten Deposit Uludağ. G. VAN DER KAADEN. Institute of Turkey, Bull. 50, Ankara, 1958. *Description of a tungsten deposit formed by high temperature solutions and localized by folding.*

El Batolito de Sauce. Pp. 52; pls. 18; figs. 2; tbls. 7. Instituto Geologico del Uruguay, Montevideo, 1956. *Geology and petrochemistry of radioactivity associated with ultrabasic rocks. In Spanish.*

Petroleum Geology (Geologiya Nefti). Pp. 9. Price, $18.00 year. The Review of Russian Geology, Columbia, S. C., 1958. *Sample copy of a Russian paper: Results of Geological Exploration for Petroleum in 1957, offered by the publishers as a service to industry and geology.*

Research and Development and Its Impact on the Economy. Pp. 21. National Science Foundation, Washington, 1958. *Bibliography of material relating to the interrelation of science and the economy.*

Idaho Bureau of Mines and Geology—Moscow, 1958.

Bull. 15. Outline of the Geology of Idaho. C. P. Ross and J. D. FORRESTER. Pp. 71; figs. 10. *Concise summary of the general geology, geologic history, geomorphology and mineral resources of Idaho.*

Pamphlet 114. Petrography, Mineralogy and Origin of Phosphate Pellets in the Phosphoria Formation. C. D. EMIGH. Pp. 57; figs. 13. *Detailed descriptions of phosphate pellets from various formations. Pellets formed by the replacement of calcium carbonate by phosphorous derived from normal sea water.*

Geological Survey of Japan—Hisamoto-chō, Kawasaki-shi, 1957–58.

Explanatory Text of the Geological Map of Japan. Iida (Kanazawa-86). K. KAWATA and N. YAMADA. Pp. 62; figs. 34; tbls. 4. *Areal geology of an igneous and metamorphic terrain in southwest Japan. Map, scale 1:50,000. English summary.*

Explanatory Text of the Geological Map of Japan. Kammuri-Jima (Kanazawa-65). O. HIROKAWA and K. KURODA. Pp. 18; figs. 2; tbls 2. *Geology of Miocene-Recent sediment volcanic sequence in central Japan. Map scale 1:50,000. English summary.*

Explanatory Text of the Geological Map of Japan. Arashimadake (Kanazawa-58). M. KAWAI, K. HIRAYAMA, and N. YAMADA. Pp. 123; figs. 39; tbls. 10. *Geology of an area in north central Japan underlain by metamorphics, Paleozoic sediments, and volcanics. Lead-zinc is mined. Map, scale 1:50,000. English summary.*

Explanatory Text of the Geological Map of Japan. Tsuyama-Tobu Okayama-44). M. KAWAR. Pp. 71; figs. 36; tbls. 11. *Areal geology of a terrain underlain by Paleozoic, Mesozoic and Cenozoic sediments and intrusives. Map, scale 1:50,000. English summary.*

Explanatory Text of the Geological Map of Japan. Tazawako (Akita-13). A. OZAWA and K. SUMI. Pp. 27; figs. 6; tbls. 3. *Areal geology of a terrain in northeast Japan underlain by Miocene to Recent sediments and lavas and by older intrusives. Map, scale 1:50,000. English summary.*

Explanatory Text of the Geological Map of Japan. Obanazawa (Okita-75). S. TOKUNAGA. Pp. 38; figs. 20; tbls. 11. *Areal geology of region in northeast Japan underlain by Miocene-Recent sediments. Coal, oil and gas are produced in the area. Map, scale 1:50,000. English summary.*

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Bulletin of the Geological Survey of Japan, Vol. 9, No. 3. Pp. 78; figs. 47. *Five papers dealing with Permian stratigraphy, ground water and engineering geology, gas and oil exploration. English abstracts.*

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Kansas Geological Survey—Lawrence, 1958.

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Economic Geology
Vol. 53, 1958, pp. 775–776

SCIENTIFIC NOTES AND NEWS

HISAKI SATO, for many years president, Mitsui Mining and Smelting Co., of Tokyo, has been made chairman of the board, and NASATOSHI KOGUCHI has succeeded him as president.

The SMITHSONIAN INSTITUTION of Washington, D. C., has opened an exhibit hall on August 1 entitled "Gems and Minerals" which includes the most extensive collection of gems on display in the Western Hemisphere. Among the collection are a 316-carat star sapphire, a 66-carat alexandrite, a 310-carat peridot, and the Shepherd diamond, a flawless canary-yellow gem of 18.3 carats. Included also is the original gold nugget that started the California gold rush. Preparation of the hall was under the supervision of curators George Switzer and Paul E. Desautels.

The U. S. MINES BUREAU is producing experimentally cerium as high-purity metal at their laboratories in Reno, Nev. Almost totally free of iron, carbon, and hydrogen, the metal made at Reno is an excellent "getter," that is, it attracts and absorbs impurities that enter other metals when they are molten. Thus, the Bureau hopes it will be useful in producing other metals in purer form. These, in turn, would be blended to make new and stronger alloys that can withstand the high temperatures encountered in rocket and missile applications.

I. G. IRVING has moved to Seattle following the shutdown of the Norwich manganese mine at Butte, Mont., which he co-managed along with his former partner, R. H. Nelson. He is now engaged in mining geologic consulting work, and is associated with Investment Exchange, 706 Securities Bldg., Seattle, Wash.

In an earlier issue it was stated that H. R. COOKE, JR. was previously "senior exploration geologist for Martin, Sykes and Associates, in South America." It should have stated that he was one of the senior geologists.

The AMERICAN GEOLOGICAL INSTITUTE at 2101 Constitution Ave., N.W., Washington 25, D. C., announces that upon request it will send copies of the First Circular for the 21st International Geological Congress in the Nordic countries in 1960, also travel literature.

The NATIONAL SCIENCE FOUNDATION has announced the establishment of a Program for Atmospheric Science. EARL G. DROESSLER joined the staff of the Foundation today as Program Director for Atmospheric Sciences. He was formerly with the Office of Science, Department of Defense. The Atmospheric Sciences Program is being established by the Foundation as a result of recommendations by the President's Committee on Weather Control and the Committee on Meteorology of the National Academy of Sciences, National Research Council. The new program will deal with meteorology, including not only the more conventional type of meteorological research, but also energy transfer processes between earth, sea, and air; turbulent flow of gaseous fluids; heat exchange processes; upper atmosphere studies (i.e., above the level of conventional meteorological work); atmospheric chemistry; and general circulation problems of the atmosphere and oceans. Attention will also be given to the field of cloud physics, especially the

physics of precipitation, where much basic research must be pursued before the possibilities of controlling or modifying weather can be evaluated.

MEREDITH E. JOHNSON, State Geologist of the New Jersey Department of Conservation and Economic Development retired after 30 years in public service. DR. KEMBLE WIDMER succeeds Mr. Johnson in the position of State Geologist and Chief of the Bureau of Geology and Topography. New Jersey's official geological work is the oldest continuous operation in America. Established in 1864, the State Geological Survey Office has had only four directors since its founding. Dr. Widmer is the fifth individual named to the post.

The U. S. BUREAU OF MINES announced that during the fiscal year 1959 four experiment stations will participiate in research on bauxite and aluminum to develop economic processes for producing aluminum from low-grade domestic materials. The research program will be carried on at Bureau Laboratories at College Park, Maryland; Rolla, Missouri; Boulder City, Nevada; and Albany, Oregon.

W. Dow HAMM, Atlantic Refining Company, Dallas, Tex., and LEWIS G. WEEKS, consulting geologist, Westport, Conn., head the slate of officer nominations submitted by the A.A.P.G. nominating committee. Other nominees include, for vice-president, ALFRED H. BELL, Ill. Geological Survey, Urbana, and LEO R. NEWFARMER, Shell Oil Company, Los Angeles, Calif.; for secretary-treasurer, GEORGE C. GROW, JR., Transcontinental Gas Pipe Line Corp., Newark, N. J., and HAROLD T. MORLEY, Pan American Petroleum Corp., Tulsa, Okla.; and for editor, DANIEL A. BUSCH, consulting geologist, Tulsa, and GROVER E. MURRAY, Louisiana State University, Baton Rouge.

The NATIONAL SCIENCE FOUNDATION announced that approximately 60,000 pages a year of key Soviet scientific and technical journals are now available in English translation to United States scientists and engineers. In releasing a compilation of translated Russian journals, the Foundation indicated that there are now in print 53 English editions of Russian journals, 4 extensive series of translated Russian abstracts of scientific papers, and 4 series of partial translations of important Russian journals, including crystallography, geochemistry, geological sciences, and petroleum geology.

DR. FRANK FITCH GROUT, Professor Emeritus of Geology at the University of Minnesota and former Director of the Minnesota Geological Survey, died Friday, August 1, 1958. Dr. Grout served the University for nearly half a century as an internationally known professor and an active research worker in Igneous Geology.

A Sub-Commission for the Metallogenetic Map of the World, under the presidency of W. D. JOHNSTON, JR., and M. P. ROUTHEIR and DR. PHILIP W. GUILD as secretaries, are asking each country to prepare a mineral map showing mineral deposits in as much detail as practicable, on the scale used for its tectonic and geologic maps, with an accompanying index indicating the magnitude of the deposits and other pertinent data. It is hoped these maps can be ready for display at Copenhagen for the 1960 International Geological Congress.

D. J. SALT, formerly managing director of Geo-Explorers Ltd., has become chief geophysicist of Ventures Ltd., Toronto.

CHARLES A. MITKE is now president of the Manila Mining Corp., Manila, P. I. The company is at present drilling what promises to become a large low-grade copper porphyry.

J. B. STONE, formerly general manager of The Fresnillo Co., Fresnillo, Mexico, has become assistant managing director of the company, replacing J. H. Ashley who resigned for reasons of health.

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